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INTRODUCTION

Keiiti Aki and William D. Stuart

We would like to begin our conference with the beautiful words of a poet who lived
in a nearby town of Carmel. One of the greatest poets of this country, Robinson Jeffers, wrote that the happiest and freest man is the scientist investigating nature, the person who is interested in things that are not human.
As you all know very well, the subject of our symposium, earthquake prediction, involves human elements at all levels. How can we be happy and free?
Robinson Jeffers also wrote that if the person is interested in human things, let him
or her regard them objectively as a very small part of the great music.
This means that we need social scientists who would study objectively the behavior of physical scientists working on earthquake prediction. As a matter of fact, one motivation of the present conference came from an article by W.H. Lambright (1985) who wrote that the Japanese decided not to issue intermediate-term predictions because intermediate-term precursors are not well founded on scientific grounds, and such predictions may cause more economic disruption and public anxiety than decision makers can handle.
The idea of intensive monitoring of short-term precursors in the area where a reliable long-term forecast of a major earthquake has been made is the wisest strategy under the current uncertainty of intermediate-term prediction. Here, short, intermediate and long-term prediction refer to the time window, up to a few weeks, few weeks to few years, and few years to few decades, respectively (Wallace et a/., 1984). It is true, however, that a reliable intermediate-term prediction will have a tremendous value to decision makers who are responsible for earthquake preparedness activities. The short-term prediction does not give sufficient time for preparation, and the long-term prediction does not give enough impact on decision makers, especially in this country where politicians have to face election every two years or so.
The intensive monitoring of short-term precursors for a specific area such as Tokai and Parkfield has another problem in that damaging earthquakes occur in broad areas outside the area where no monitoring works are done. There is a need for monitoring of intermediate-term precursors over broad areas.

Ideal case of earthquake prediction

As mentioned earlier, one of the tasks of social scientists in earthquake prediction is to objectively investigate the behavior of physical scientists working on earthquake prediction. Under this circumstance, we must carefully define our task if we still want to be happy and free.
We would like to define the task of a physical scientist in earthquake prediction as estimating objectively the probability of occurrence of an earthquake with a specified magnitude, place, and time window under the condition that a particular set of precursory data was observed.
The ideal prediction in this form would provide the probability rate of 1 earthquake/t per unit time at time t before the occurrence of the target earthquake as shown in Fig. 1. Namely, 100 years before the earthquake it will predict the probability rate of 1 per

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100 years, and 1 day before the earthquake, it will predict the probability rate of 1 per day.
In Fig. 1, we also show schematically the state-of-the-art in earthquake prediction. The horizontal line at the bottom corresponds to the long-term average rate of occurrence which can be estimated from an earthquake catalog and paleoseismological data. The curve marked as 'gap theory' indicates roughly an order of magnitude increase in probability from the long-term average based on the seismic gap along a plate boundary. According to S.P. Nishenko (1988), 13 earthquakes have been predicted successfully by this method. The vertical line marked as 'SCAN' (Seismic Computerized Alert Network) is not a true prediction but an early warning by detecting the occurrence of a major earthquake. This type of warning has been in operation for many years by the Japan National Railway. The Parkfield experiment (Bakun et a/., 1986) is intended to achieve a short-term prediction, in the manner of the Tokai earthquake monitoring, based on seismic, creep, continuous strain, and geodetic observations. We note that there is a great gap between the ideal case and the intended curve in the intermediate-term range.
The curve marked as 'Haicheng' was constructed by Cao and Aki (1983) using Utsu (1979) method for assigning probability gain to various precursors of the Haicheng earthquake of 1975, and is by far the closest to the ideal case. In constructing the curve it was assumed that precursors with distinctly different precursor times are independent. The probability gain was estimated from scanty information on the precursor time and the success rate for each precursor. A difficult question is whether this Haicheng curve expresses the physical reality in the earth or the wishful thinking of precursor observers. There was a strong political pressure at the time in China to encourage reporting of precursory phenomena. The pressure must have helped to make a complete collection of precursory data by the participation of the masses, but also must have induced reporting noise as a precursory signal.
Thus, we must assign a great range of uncertainty to each curve of probability of earthquake occurrence. The range of uncertainty represents the range of our ideas and hypotheses about the physical process of earthquake occurrence. It is important to convey this range of uncertainty to the decision makers so that the decision makers can judge the state-of-the-art of earthquake prediction properly.

The. purpose, of the, present symposium

Intermediate-term prediction may be more difficult than the short-term prediction also from physical point of view because the precursory signal reflecting the state of stress may be weaker a long time before than immediately before the time of earthquake occurrence. We feel, however, that there are some recent encouraging developments both in the observational and physical bases of intermediate-term prediction.
The source of optimism of one of us (K.A.) on the observational basis is the temporal change of the quality factor of earth's crust as measured from the coda waves of small local earthquakes. One advantage of coda Q as a precursor is that the data are available for a uniform coverage of many seismic regions, and a systematic testing of the method is possible without additional measurements. A critical review of the reports on coda Q precursor will be given by H. Sato in this symposium.

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The source of optimism on the physical basis shared by both of us is the ability of slip-weakening type friction laws in explaining various aspects of earthquake phenomena including the recurrence behavior of characteristic earthquakes, frequency magnitude for small earthquakes, scaling law of seismic spectra for small earthquakes, fmax of large earthquakes, and the dependence of seismicity quiescence precursor on the cut-off magnitude. The last mentioned may be most relevant to the subject of this symposium and will be discussed by T. Cao in his presentation in this symposium. The key parameter of the friction law is the critical weakening slip which appears to control a variety of crucial phenomena.
We know some of you are as optimistic as we are but for different reasons from ours. We know also some of you are pessimistic about the sources of our optimism. We think we have a healthy mix of optimism and pessimism to promote a happy and free science for the next three and a half days.
This meeting will be followed by a meeting of the U.S. National Earthquake Prediction Evaluation Council. We are requested by the chairman of the council, L.R. Sykes of Columbia University, to answer whether some of the intermediate-term precursors such as the seismicity quiescence are ready enough for use in the routine operational earthquake prediction. It would be a great accomplishment if we can do that. This question of 'ready enough' was also discussed by W.H. Lambright in the article mentioned earlier, and clearly involves human elements, such as the need of local community. Again, we feel that all physical scientists can do is to give an objective estimate of probability of occurrence of the target earthquake together with the range of uncertainties about the probability estimate. Decision makers should proceed from there with the help of social scientists.

Acknowledgements

The idea of having this conference arose during a meeting of the National Earthquake Prediction Evaluation Council. The convenors thank Professor Lynn Sykes, Chairman, and the other members of NEPEC for encouragement and for endorsing the conference. The convenors and participants are grateful to Wanda Seiders and Barbara Charonnat for their assistance in organizing.

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Probability Of Earthquake Occurrence
In The Next Minute

10

10

10

- 2

- 3

10-4

10

10

- 6

10

- 7

10

- 8 Long-term Average

I

lOOyr

/t^ A fo A A

lOyr. lyr 1 mo. Iday Ihr
TIME BEFORE EARTHQUAKE

1 min.

FIGURE 1

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STUDY ON INTERMEDIATE-TERM EARTHQUAKE PREDICTION

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Abstract

This paper proposes that the upper limit of intermediate-term prediction should be extended to about 20 years from several years, and that intermediate-term prediction should be divided into two stages: an early intermediate-term stage (>2-S years) and a late intermediate stage. Seismicity features in both stages are closely related and are difficult to distinguish. The pre-earthquake episode of about 20 years, during which a group of strong earthquakes occurs, is the main focus of early intermediate-term prediction. The concentrating process of small earthquakes along belts and from far-field to near-field 2- 8 years before strong earthquakes, accompanied by several anomalies in geophysical and geological fields, is the main focus of late intermediate prediction. Precursors before nine strong earthquakes (Ms > 7) during 1966-1976 are described in the paper. Finally, four mechanical models are suggested for possible use in intermediate-term prediction.

Introduction

Intermediate-term earthquake prediction lies between long-term and short-term prediction in time-scale and has relatively independent research methods and aims. Based on experiments and results of Chinese earthquake prediction studies, this paper discusses the time scales, important phenomena, and physical models relevant to intermediate-term earthquake prediction.

Precursory Stages of Earthquake Prediction

Table 1 gives six schemes for classifying precursors according to their time before an earthquake.(^ By comparing these schemes and considering the present knowledge about preparatory processes of earthquakes, the state of data, and the convenience for working programs and international exchange, we think that the sixth scheme in Table 1 is the best. This scheme suggests that long-term (>20 years), intermediate-term (20 years to several months), and short-term (< several months) prediction may be possible.
Most workers agree that a precursor occurring several decades before an earthquake would be a long-term precursor. The research methods of long-term prediction all depend on recurrence of strong earthquakes, revealing long-term regularity of strong earthquakes at a location, a belt, and a region. Research results have been used for earthquake zoning and evaluating of seismic risk.
Most workers also agree that precursors occurring several months or less before
a strong earthquake are short-term precursors because abrupt and rapidly-changing

anomalies of geophysical and geochemical fields before strong earthquakes, in features and time scale, are very similar to each other and also to certain experimental and theoretical results. Moreover, it appears that the more imminent a strong earthquake is, the more abundant the abrupt anomalies are, and the higher the ratio of different macroscopic anomalies, including unusual behavior of animals and climatic anomalies. People generally expect to distinguish an imminent-term prediction several days before a strong earthquake from the short-term stage, but several peaks of abrupt anomalies often appear before a strong earthquake. Thus, in most cases, it is difficult to determine definitely the start of the imminent stage.
According to the different schemes in Table 1, there is some divergence of views about the time range and content of intermediate-term prediction. The intermediate-term is situated in the middle time scale, and its physicaland practical basis and practical meaning also show interim character. The data, which will be discussed below, indicate that in the period from several years to several decades before a strong earthquake, there are regular seismicity pattterns which are closely related to precursors occurring in the period from years to months. Every kind of slowly-changing geophysical precursor appears mainly 2-3 years before a strong earthquake. Therefore, these two time scales may be represented by two stages: the early intermediate stage, and the late intermediate stage. But the time limit dividing both stages is not very stationary, and phenomena belonging to both stages are difficult to distinguish. Therefore, it is better to merge both stages into an intermediate-term period. Research on intermediate-term prediction, besides being used directly for evaluating seismic risk for engineering studies, can be applied in two ways:
(1) to check and correct results of long-term prediction, and (2) to decrease the ambiguity of short-term predictions of place and earthquake magnitude.
Since 1972, formal intermediate-term earthquake prediction has been carried out in China. During 1983-1985, experimental research on regional intermediate-term prediction was carried out in North China and the Yunnan-Sichuan seismic area to infer forthcoming focal areas before the year 2000 A.D. This work in our country trys to combine long-term and short-term methods into a new procedure.

Seismic Period, Seismic Episode, and Heterogeneous Distribution
of Earthquakes in Time and Space

Earthquakes occur often in China, and strong earthquakes (Ma > 6) occur in nearly every province of China. In China there are nearly 70 places where large earthquakes (Ma > 7) have occurred and 250 places where strong earthquakes (M3 > 6) have occurred; earthquake recurrence times and magnitudes in these places often change. (2) Of the nine large earthquakes (Ma > 7) occurring in China during 1966-1976, six large earthquakes occurred in places where no large earthquakes (Ms > 7) occurred before.
To probe the cause of the variation of earthquake recurrence times, records of historical earthquakes from several hundred to 2,000 years have been widely studied in China, and seismic periods of about 300 years and seismic episodes of about 2 or 3 decades have been found.(3' Some of the research results bearing on seismic episodes related to intermediate-term prediction will now be discussed.

Earthquake activity in North China since 1400 may be divided into two active seismic periods:^3) (1) 1484-1730, and (2) 1815-present. No strong earthquakes (Ma > 6) occurred in the 85 year interval between these two periods, and present seismicity belongs to the second period which would end before 2030 if the two periods are the same length. Each seismic period may be divided into several active seismic episodes (Fig. 1), but the earthquake distributions in the episodes and their relation to seismotectonics differ. During this century China and adjacent areas have suffered four active seismic episodes^ 4) (Fig. 2), and they are now facing a fifth active seismic episode. From Fig. 2b, we can see that earthquakes (Ms > 7) of each episode occurred in one or two regions or seismotectonic belts. Fig. 2a shows that periods of earthquake recurrence at 20 year intervals exist in the entire country, but such regularity of recurrence does not apply to individual seismotectonic belts or locations of strong earthquakes.
From regularities of earthquake distribution and epicenter shift in each seismic episode,
we can see that between seismic episodes there are obvious differences which reflect the variation of the regional stress field for each seismic episode. Such variations reflect the micro-dynamic state of the regional stress field, and several strong earthquakes in each seismic episode occur as the successive results of the unified stress field in each episode. Therefore, variations of recurrence interval and earthquake magnitude in places with frequent strong earthquakes are controlled by recurrence regularity of seismic episodes and a micro-dynamic state of stress field in each seismic episode.
In conclusion, recurrent phenomena related to earthquakes are not only determined
by local crustal conditions, but are also related closely to the micro-dynamic state of the regional stress field. These recurrent phenomena appear as different levels of rhythm of earthquake swarms, such as seismic period and episode. It is from such observations that principles and methods of long-term prediction are inferred.

Seismicity Concentration toward the Source Area During the Intermediate-Term Stage

Nine large earthquakes with Ms > 7 occurred in succession from 1966 to 1976 in North China and Chuan Dian (the Yunnan Province and the west part of Sichuan Province) area, which are the main active areas of the fourth active episode of strong earthquakes since the beginning of this century (Fig. 3).
Fig. 4 shows the seismicity concentration from the far-field to the near-field before eight large earthquakes.(5) The aftershock areas of the main shocks are marked by dotted lines, and the near-field areas of seismicity concentration by broken lines. The areas outside the broken lines are called the far-field areas. The long axes of near-field areas are two or three times longer than those of the aftershock areas. The seismicity concentrations in the near-field areas last about 2 or 3 years.
To examine the beginning time and distribution of the seismicity increase in the far-field area, seismicity on a larger scale has been studied. The maps of epicenter distribution at different stages (1961-1976.7) before the Tangshan earthquake in North China (114°E-120°E, 38°N-41°N) are shown in Fig. 5.^6 ^ The main features of observations in Fig. 5 are the following:

(a) From 1961 to 1963 (Fig. 5a), seismicity was dispersed over the entire area.
(b) During the period of 1964-1966 (Fig. 5b) seven shocks (M > 4) occurred, but six of them were concentrated in a belt with NW orientation along Tienjin-Beijing-Huailai; several strong shocks occurred historically in this important seismic belt.

(c) From 1967 to 1969 (Fig. 5c), the seismicity was more discrete, but was still mainly distributed in the above NW belt. A moderate earthquake (M = 5.7) occurred at the NW end of the belt, and the Bohai shock (M3 = 7.4, July 18, 1969) occurred at the SE end of the belt.

(d) From 1970 to 1972 (Fig. 5d), seismic activity increased in the NW belt, six earthquakes with M > 4.0 took place, and there were few small shocks; thus the b-value of the whole belt decreased. During the 9 years from 1964 to 1972, sixteen earthquakes with magnitude 4 or more occurred in the above NW belt. These sixteen earthquakes made up eighty percent of all earthquakes M > 4.0 in the whole area. Notice that there is a gap in the Tangshan-Tienjin-Beijing region of the seismicity belt. The gap coincides with the near-field region of the Tangshan earthquake, and so increased seismicity in the far-field before the Tangshan earthquake was a stage in the process of seismicity concentration in the Beijing-Tienjin seismic belt.
(e) Fig. 5e shows the case of seismicity concentration from 1973-1976.6 in the near-field of the Tangshan earthquake. In this stage, there were seven earthquakes M > 4.0 distributed with NE orientation in the above gap. This was the late seismicity concentration of the intermediate stage in the near-field. In other words, it was
an early foreshock phenomena in the near-field. It was followed by the Tangshan earthquake (M3 = 7.8) near the intersection of the NE and NW seismicity belts.

Three main points can be inferred from the above data: (1) the seismicity concen-tration toward the near-field during the intermediate-term stage lasting 3 lfa years was a continuation of early intermediate-term (9 years) seismicity development within the dense NW seismicity belt; thus the early and late intermediate-term stages were closely related;
(2) the Bohai earthquake (M3 = 7.4) of July, 1969, occurred in the other gap of the NW belt, and the Tangshan and Bohai earthquakes were related mechanically, both having intermediate-term seismic anomalies; (3) from the point of view of seismicity in the whole of North China, the Xingtai earthquake (March, 1966), the Bohai earthquake (July, 1969), the Haicheng earthquake (February, 1975), and the Tangshan earthquake (July, 1976) oc-curred in the background of the same episode of regional seismicity. Another example is the succession of five strong shocks from 1970 to 1976 in the Chuen-Dian area.

Intermediate-Term Precursors of Strong Earthquakes

There were also other seismicity and geophysical anomalies before the nine large earthquakes discussed above. Figs. 6 and 7 are examples before the Haicheng, Tangshan, and other earthquakes, respectively.^7' 8 ^ The curves enclosing observations in Fig. 8 indicate the time-space distribution of precursor anomalies. All axes are logarithmic. Fig. 9 is a "window" of inferred stress field in North China. It shows apparent anomalies before several strong earthquakes.

The following four points summarize anomalies in Figs. 6-9 starting at least 2 to 3 years before a strong shock with magnitude 7 or more.

(1) Several kinds of seismicity anomalies are observed in the first 2 or 3 years before the occurrence of a large earthquake. They mainly include rapid increase or decrease of shock frequency and released energy; they are expressed in space as a seismicity band
or gap. At the same time, anomalies of regional land deformation appear. All these anomalies are related to the development of the seismic episode.
(2) During the period of 2 or 3 years before the occurrence of a large earthquake, seismicity concentrates in the near-field area. At the same time, there are many anomalies outside the near-field including land deformation, water level, chemistry of underground water, and earth resistivity (Fig. 10), which are related to tectonic strain and displacement in the shallow part of the crust. In addition, geomagnetism and gravity anomalies are observed in a few cases.
(3) Approximately a half year prior to a strong earthquake, anomalies change nearly simultaneously in the near- and far-field. At the same time, new anomalies such as level and radon content of underground water and earth tilt occur in places without previous anomalies (Fig. 8b).
(4) Within a week to a month prior to a strong shock, rapid and sudden anomalies appeared in a large area (Fig. 11). The anomalies show a concentration toward the near-field.

The above four stages are equivalent to early intermediate-term, late intermediate-term, short-term, and imminent-term, respectively. In the ideal case, these four stages can be divided, but in the general case, the former two stages can be joined to intermediate-term, and the latter two to short-term. All of the above mentioned anomalies might be observed under favorable conditions, but there are three main problems: first, how to distinguish the true from the false anomalies; second, anomalous intermediate-term precursors are not a unique indicator of regional tectonic variation, and are reversible; and, third, in the preparatory process of several strong shocks close to each other, there are very complicated effects.
The distance relation between epicenters and anomalies is complicated in our country.
In general, there may be two cases, as in Fig. 12. Case A indicates a single epicenter, and intermediate-term anomalies appear not only in its near-field, but also in its far-field. Case B indicates that an intermediate-term anomaly may respond to several large earthquakes which successively occur in a region. Thus, except for the focal action on intermediate-term anomalies, we suggest that intermediate-term anomalies reflect regional stress and strain fields, and that large earthquakes occurring successively are regional seismic phenomena.

Discussion of Models

Multiple Stress Concentration Model

To explain the mechanical relation among strong shocks which occurred successively within the same seismic region, the author proposed a multiple stress concentration model in 1981 (9), as shown in Fig. 13. Under the remote stress field, the stress concentrations

occur where relative block motion or fault slip is impeded. Unlocking points are sites of earthquakes and allow energy release of the stress field and change in stress distribution. The elastic energy E of the stress field can be expressed approximately by the sum of the product of stress (at) and strain (et-) at the locked points with different areas,

The variation in elastic energy of the stress field with time depends on the difference between the time change in regional force F(t) and the process of earthquake energy release (-E"o)- The unlocking order of the locked points depends on the regional force, block geometries, and the structure and strength of locked points.

Viscoelastic Element Model

To explain the successive occurrence of strong earthquakes in Japan, Mogi^ 10' proposed a spring and slider model whose phenomenological elements are shown in Fig. 14a. To explain why the time intervals between earthquakes in a group of strong shocks shorten exponentially, Chang Guomin et al. presented the viscoelastic model in Fig. 14b and 14c(n). This model supposes that the rupture strengths of different elastic elements are the same and that the shortness depends on progressive risk of the total stress level in the model (Fig. 14d). In fact, the structures and strengths of locked points are different, and the rate of stress increase is not the same for each. When the number of locked points decreases, the rate of stress increase on the remaining locked points would increase.

Weakening Source Inclusion Model
To explain seismicity concentration toward the source area, Fu Zhengxiang (1984)(5 ' modified the elastic inclusion model for an earthquake source. When the change in stress and strain is at stage QA' shown in Fig. 15c, the rigidity and stress of the inclusion are higher than those of the elastic surroundings. When it is at stages AB and AB' after increase of the regional stress level, the inclusion weakens and its rigidity decreases. During this stage precursory seismicity will take place mainly in the inclusion. Later, when the inclusion stress reaches the peak value, the slope decreases, and the inclusion is strain weakening. When the negative slopes of the two curves become equal (Eshelby line), instability occurs, producing the equivalent of an earthquake. The earthquake inclusion model can help understand migration and concentration to the future source area before a large shock.

Fault Creep Model

By inverting earth deformation data measured near the source areas of the Xingtai, Haicheng, and Tangshan earthquakes, Chen Yuntai^ 12) and Zhang Yinghen et alS 13 ^ proposed a fault creep model of the source area and described the fault creep process accompanying the increase of regional small earthquakes before a strong earthquake. The concentration of small earthquakes for the Beijing-Tienjin NW seismotectonic belt before the Bohai and Tangshan earthquakes suggests that fault creep occurs before strong earthquakes in China. Savage^ 14) and Fu Zhengxian^ 15) discussed the depth of a creeping

fault, its relation to the middle and lower crust, and upward migration of creep. The model needs additional testing.
In conclusion, the multiple stress concentration model and the viscoelastic model may help explain repeated occurrences of strong earthquakes and their mechanical and historical relations over one or two decades. The models may also provide a theoretical basis for early intermediate-term prediction of a strong earthquake group. On the other hand, a combination of the inclusion model and the creep model may help explain anomalous phenomena during the late intermediate-term before a strong earthquake.

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^\<u-u

^§\
time >v

- 100 years

- several years

several months

- several days

before

shock

after

- several days

- several months

- several years

1

earth deformation

Mecherlkor , J.A.
1968

o(deformation
slow deformation

/3 deformation

20-25 years

4-5 years

r sudden
def ormat ion

Fujli Yolchlro
1974

/^obvious
deformation

(9A unstable
dcformat ion

A Slip
* one year

I
tens minuts

r dislocation

r. dislocation

r. continue

slip

Inverse
recovery

2

examples of quakes

Ma Zongjln
1972

Super long-term
seismic period

lont-term

seismic episode
(20_+ years)

intermediate -term

slow precusor

short-term
rnpld precusor

1 nn tent

sudden precusor

d Islocat Ion

af tershock-
dis location

adjustment

3

experiment

Scholz.C.H.
1982

T-strain
slow accumlation

TI

mainly dilatec

111 ma Inly
soaked

IV rapid
deformation

V dislocation

a f ershock
adjustment

theory

m'iachking(V
1975

I-straln
uniform
rupture

II Micro-
rupture
avalanche

III
concentra-
tion of
crack

IV chief
rupture
onloadlng

4

example of quakes

Rikitake, T
1976

long-term
strain accumalatlon

(decades)

intermediate-term
dilatation

6-7 years

short-term
2-1 years

imminent

chief rupture
p re-developed

5

antiselsmlc request

Mumrpe.T
1978

long-term
10 years

intermediate!- term

6-10 years

short-term
6 months

several. days
subshort-term

6

summarization

Ma Zongjin
1986

long-term

intermediate -term

early

late-

Jhoirt-tenn

shock time

strong
. aftershock

adjustment

Table 1

Comparison of Earthquake Prediction Stages

M

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K-a90 B- 19 76 H

""1900
10 20 30

O A/=7.0 7.9 O A^=8.0 8.4 Q M>t.5 £3 major active region

Fig. 2. Seismicity patterns of four seismic episodes in China. (a) Four active seismic episodes in Chinese mainland and its neighboring regions, (b) Spatial distribution of the four active seismic episodes in Chinese mainland and its neighboring regions during this century.

HAICHENG NU7.3

40

30

Z 20

10

0

> 1-75 I ,

1.0

0.5

1S72 73 74 75

TIME (yeor)

Fig 6 a

Change* in then intermediate-tern anoe-alie* before Haicheng earthquake. A) change in magnitude with tine in Liaoaing Province; B) avail earthquake frequency in south Liaoning Province; C) ratio of wave velo-city; D)fault displacement at Jin Can short levelling atation (A 190 km); F) ground tilt at fingkou station (A » IB ka).

TANGSHAN Mj = 7.8

1.5

b 1.0

0.5

174
1.64
L54

B

1968 69 70 71 72 73 74

TIME (yeor)

75 76 77

FigGb
Changes in intermediate-term anomalies
before the Tangshan earthquake. A) eerth-quake frequency (M >. 4.0) in the Tianjin-Tangshan region; B) b-valu« in Tangshan region; C) ratio of wave velocity in the Tianjin-Tangshan region; D) Level of under-ground water at Tangshan well; E) resisti-vity at Changli station (A 70 ka); F) average change in radon content of under-ground water in Beijing-Tianjin-Tangshan region; C) fault movement at Tahuishag sta-tion (A* 200 ka); B) fault Movement at Hiukouyu station (A - 180 ka).

Fig. 6. Intermediate-term anomalies before the Haicheng and Tangshan earthquakes.

16

levelling
9- Jin Xian

(mm)

Haiclieng7.J

*

levelling Xtaguan
170km

2 (mm)

levelling
2 (mm

Songpanr.2
Songpan 1

(b)

-9 -8 -7 -0 -5 -« -3 -2 -1 0 (yr) -S -4 -5 -J -l 0 (yr)

Fig. 7. Comparison among intermediate-term anomalies before several strong earth-quakes, (a) Change in leveling, (b) Change in number of regional seismicity (Mi > 2.0), (c) Change in b-value. (1) South Liaoning area, (2) Longling area
and surrounding, (3) Tianjin-Tangshan area, (4) Longmenshan area.

(day)

100

V Rn

(day)

1000

100

10 100

(t) Tangshan earthquake

1000 km 10 loo (b) llaicheng earthquake

looo km

LI :sea level;
LI : short levelling ;
Lj : tilt;
Si : seism icily ;
Sj : seism icily bell;
Sj : Seisin icily .window
(Hongshan);
84 : b value;
PH : resistivity ;
Z: earth magnetism;
C: atmosphere phenomena;
A: animal behavior;
0: earth light i
G: gravily ;
Rn : radon content ;
W: underground water.

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Fig. 8. Time-space distribution of strong earthquake precursors.

O/V/L4 4.9 O-5 5.9

O epicenters from 1966 to 1972.8
Sepicenters from 1972.9 to 1976.6
Swater level clevelling A Radon
dearth magnetism © resistivity
9 30 60 90 120km

Fig. 10. Space distribution of intermediate-term anomalies before the Tangshan earth-quake.

Shenyang'

hohhot

ATaiyuan

"Cangzhou
A Shijiazhuang

"~^"® Zhengzhou

A Radon content
« water level
O resistivity
" oil well
0. 60 120 180km

Fig. 11. Space distribution of sudden anomalies before the Tangshan earthquake.

21

..surrounding

inclusion

surrounding
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surrounding
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D'insUbilily
inclusion

Fig. 15. Model of weakening source inclusion, (a) Constitutive relation of inclusion and surrounding, (b) Rice's model of seismic gap. (c) Model of seismicity concentration and migration.

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REPORTED INTERMEDIATE-TERM EARTHQUAKE PRECURSORS IN THE IZU PENINSULA, JAPAN

by

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Abstract

Reported intermediate-term precursors of the 1978 and 1980
Izu earthquakes, crustal uplift, earthquake swarms, volumetric strain anomalies and anomalies in the temperature, level, and radon content of well water are re-examined based on the data as long as available. Especially, it is attempted to judge whether a reported anomaly is significantly different from changes which do not precede large earthquakes.

The most significant ones are an anomaly in the radon concentration of well water and that of water temperature at a different well both of which took place in December, 1977, more than a month preceding the 1978 event. However, the reported similar pattern of precursory change in various sites is found to be groundless. The various data may suggest a decrease in upflowing water and therfore crustal expansion in the norther part of the Izu peninsula and contraction in the southern part about a month before the 1978 Izu-Oshima earthquake, which is consistent with the mainshock mechanism.

The localized crustal uplift of the eastern coast,
and earthquake swarms took place not only before the 1978 and 1980 events, but also before the M=7.3 North-Izu earthquake of 1930. But the precise location, dimensions, and lead time before the following event were different for each event and there is no clear quantitative empirical correlation between the lead time or diameter of uplift and the earthquake magnitude. However, the similar maximum uplift rates suggest a common physical mechanism.

It is most likely that each of a large earthquake, episodic uplift, and an earthquake swarm is resulted from some common but unkown tectonic process. The source mechanisms of the large earthquakes and swarm earthquakes are generally consistent with the tectonic regime in this region, where a bending of the Philippine Sea plate before its subduction plays an important role. It is suggested that earthquake swarms are caused by a sudden increase of pore pressure associated with dike injection derived from intruded magma which causes uplift.

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Introduction

The Izu peninsula lying 100km southwest of Tokyo constitutes
an ideal natural laboratory for earthquake prediction research (Fig. 1). The crustal uplift and earthquake swarm activity have been occurring intermittently for more than a decade, the latest major episode of swarm at the time of writing, taking place in October 1986. Three large damaging earthquakes took place during this period: the M=6.9 Izu-Oki earthquake of May 7, 1974, the M=7.0 Izu-Oshima earthquake of Janauary 14, 1978, and the M=6.7 East Izu (or Izu-Toho-Oki) earthquake of June 29, 1980, (The magnitudes are those given by the Japan Meteorological Agency unless otherwise sated. The local time, JST = GMT + 9hours is used.) Similar activity was observed in the peninsula in 1930s involving the M=7.3 North Izu earthquake of November 26, 1930. The activity ended after the M=5.5 South Izu earthquake of 1934 and the quiescence lasted forty years till the present enhanced activity started in 1974. The inflation of the northeastern part of the peninsula which took place in 1930 to 1932 partially collapsed during this quiescent period.

Fairly dense observational networks are located partly because of an impending Tokai earthquake anticipated in the Suruga bay, west of the peninsula (Fig. 1). According to the recent compilation of reports on precursory phenomena by Hamada(1986), 34 anomalies of various kinds are reported to have taken place prior to the 1978 and the 1980 events, among 83 prior to 10 damaging Japanese earthquakes in the past decade. Various intermediate-term precursors are reported for the 1978 Izu-Oshima earthquake. Most of them were found after the earth-quake because only a few kinds of data were telemetered at that time. Reported precursors are: crustal uplift, earthquake swarms, volumetric strain anomalies, and anomalies in the temperature, level, and radon content of well water. They were summarized by Mogi(1981, 1986), Wakita(1981), and Wakita et al.(1985) and are shown in Figures 1 to 3. In contrast with those reported for the 1978 event, only a few anomalies are reported prior to the 1980 event.

Since several kinds of observation have been carried out continuously by the time of writing, it would be possible to make a statistical judgement on whether a reported anomaly is significantly different from changes which do not precede large earthquakes. We will re-examine reported intermediate-term anomalies based on the data as long as available. This paper consists of two parts. In the first part, reported intermediate-term precursory changes in the temperature, level, and radon content of well water and strain will be examined. Especially, similarity of these changes pointed out by Wakita(1981) will be

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discussed. In the second part, synchronized activity of crustal uplift with an earthquake swarm will be described. It is suggested that each of a large earthquake, episodic crustal upheaval, and an earthquake swarm represents only a part of some unknown tectonic process and that no causal relationship exists between a large earthquake and episodic crustal uplift associated with earthquake swarms.

Reported similarity of intermediate-term precursors
before the 1978 Izu-Oshima earthquake

Figure 2 shows reported precursory changes of the 1978 Izu-Oshima earthquake complied by Wakita(1981) and Wakiata et al.(1985), all normalized to the same time scale. Wakita(1981) pointed out that the patterns and precursory time of these changes are similar to each other. Furthermore, Wakita et al.(1985) suggested that simultaneous strain chages of the order
- 6

of about 10 took place throughout the Izu peninsula prior to the Izu-Oshima earthquake.

In the following, each reported anomaly will be re-examined based on the data as long as available. The most significant changes are anomalies of radon content (Fig.2-a) and water temperature (Fig.2-b). However, apparent similarity of the patterns shown in the figure turns out to be due to an artefact. Also the stability of the water level (Fig.2-c) shown in the beginning of the figure is due to an artefact. The strain changes (Fig. 2-d) appear to be statistically significant, but similar and more distinct change of reversed box-car type was observed at other two staions without a subsequent large earthquake. There is not long enough data to assess the significance of the changes shown in Fig.2-e and f. The occurrence of earthquake swarms (Fig. 2-g) may be regarded as the intermediate-term precursor (Wakita, 1981), but it would be difficult to distinguish them from other numerous earthquake swarms which were not followed by a large earthquake.

Radon anomaly at SKE

The most significant intermediate-term anomaly prior to the Izu-Oshima earthquake of Jan 14, 1978 would be that in the radon concentration of groundwater at station SKE in the middle part of the Izu peninsula (point a in Fig. 1 and Fig. 2-a). The continuous observation for almost a decade has been carried out by Wakita and his colleagues at an artesian well with a depth of 350m (Wakita et al., 1980a, Wakita, 1981, Wakita et al,. 1985, 1986).

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Figure 2-a shows bi-hourly readings on the original record from September 1977 to February 1978. The absence of the data between October 24 and December 15, 1977 is due to an imperfect connection of the signal cable to the recorder (Wakita et al., 1980a). Variations starting from the middle of October and the rapid change which took place on January 9, 1978 were interpreted as the intermediate- and short-term precursors to the Izu-Oshima earthquake, respectively (Wakita et al., 1980a, Wakita, 1981). The radon concentraion started to .decrease in the middle of October and fluctuated rapidly untill the end of December. Then it stayed at the lower level more than a week. A sudden drop took place on January 8 S After staying at the minimum level for 8 hours, the radon concentration rapidly increased on January 9. No significant co-seismic change was observed, but apparent enhancement lasting about a week was observed after the Izu-Oshima earthquake.

Fig. 4 shows almost 8 year-record of the radon observation
at the same site (Wakita et al., 1986). An annual variation, high in the summer and low in the winter, can be seen in most of the years, together with'a trend with lower frequencies. The most significant change is the rapid change on January 8 and 9, 1978. The radon concentration became lowest in February to March in 1979, 1980, 1981, and 1982, but it became lowest on January 9 in 1978. The rate of decrease in radon concentration staring from mid-October of 1977 (Fig. 2-a) is roughly twice that in normal autumn to winter period. However, a similar high rate of decrease was observed at the beginning of 1981, which was not followed by a large earthquake. Perhaps the most significant intermediate-term change is a rapid decrease in radon concentration starting around December 10, 1977. Except for a short-term sudden decrease taking place several days before the Izu-Oshima earthquake, a similar high rate of decrease has not been observed, though a similar rate of increase has been observed. The decrease in radon concentration starting around December 10 at station SKE appears to coincide with a decrease in water temperature (Fig. 2-b) at I-4 which is located 8km from SKE.

The M=6.7 East Izu earthquake of June 29, 1980 took place at almost the same distance from staion SKE as the M=7.0 Izu-Oshima earthquake of January 14, 1978 (Fig. 1). However, no significant anomaly was observed for this event.

Water temperature anomaly at 1-4

Nagai et al. (1979) reported an anomalous change in water temperature of an artesian well (numbered 1-4) with a depth of 500m in the middle part of the Izu peninsula (point b in Fig. 1). Fig. 2-b shows readings of water temperature flowing out from the

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well, measured once a day at the same time of day at the ground level. Fig. 5 shows an almost two-year record of the observation at the same site and a schematic diagram showing a configuraion of the top part of the well and the measuring point. The largest change except for the one precursory to the Izu-Oshima earthquake is a roughly 5°C change in September, 1976 and the record shows a good stability of the temperature. Although the available record length is much shorter than that of the radon concentra-tion at SKE, the observed anomaly prior to the 1978 event appears to be significant.

However, according to Nagai et al. (1979), the temperature decrease starting from December, 1977 and subsequent changes are due to a decrease in the water pressure of the artesian well and a change in the condition of outflowing. This indicates that a simple comparison of Fig. 2-a with b is misleading. In the following, some details of their argument will be described.

Because of high water pressure of this self-spouting hot spring, a 5m high tube is attached to the top of the well and outflowing water is pooled in a tank shown in Fig. 5-b. The water temperature was measured at point A shown in the figure after opening cock B for measurement. Untill the end of Novemver 1977, the water head was higher than 5m and the temperature showed good stability. Nagai et al.(1979) attribute the lowering of temperature starting from December 10 and subsequent rapid fluctuations to that the water head did not always reach to the top of the tube because of a rapid decrease in water pressure. When it did not reach to the top, the outflowing stopped and the water in the tube was cooled by the air.

The flow rate slowly decreased from 200 1/min in April,
1977 to 180 1/min in December, 1977. In accordance with it, the water temperature slowly decreased (Fig. 5). The rate dropped to 150 1/min after the Izu-Oshima earthquake. In February it further decreased to 80 1/min and finally self-spouting stopped in March 1978. After the earthquake, cock B was kept open. Thus the water tempeature became suddenly increased.

Most of the temperature changes shown in Fig.2-b are due to
an artefact and special care must be taken to compare Fig.2-b with others. However, the inferred decrease of water pressure appears to coincide with the intermediate-term radon anomaly mentioned in the preceding section. The both observations may suggest a decrease of water outflow taking place more than a month preceding the 1978 event.

Water level anomaly at 1-6

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Figure 6 shows 1.5-year record of water level at an artesian well (numberd 1-6) with a depth of 500m in the middle part of the Izu peninsula (point c in Fig. 1). The water level is continuous-ly measured every 5 minutes together with barometric pressure and precipitaion by Kishi(1979). Five-day average values of water level and total precipitaion in five days are shown in the figure. The blow-up of the changes in water level associated with the Izu-Oshima earthquake is shown in Fig. 2-c.

The water level appears to be quite stable at 1-6 from the figure, but this is mainly due to a fact that a float for measurement was stuck to the well wall (Kishi, 1979). The varia-tion of the measured water level decreased in February 1977 and small variations associated with changes in barometric pressure became undetected in April. The float was cleaned in early October and then small amplitude variations became recorded associated with barometric pressure change. Thus the stability of the water level shown in the beginning of Fig. 2-c is due to an artefact. The water level started to decrease in early December, more than a month before the Izu-Oshima earthquake. The level returned to the former one just before the event, and then dropped at the time of earthquake.

The significance of the lowering of water level in December,
a supposed intermediate-term precursor, is uncertain. The available real record of water level is short because most part of the record shown in Fig. 6 is affected by a stuck float. Lowering of water level similar to that in December 1977 can be seen in December 1976 and in January 1977.

Volumetric strain anomaly

Yamagishi et al. (1978) reported an anomalous volumetric strain change observed at station NGT (point d in Fig. 1) prior to the Izu-Oshima earthquake (Fig. 2-d). Relatively high rate of contraction started on December 4, 1977. On January 11, 1978, three days before the occurrence of the Izu-Oshima earthquake, the sense of the volumetric strain change reversed and relatively high rate of extension started.

Wakita et al. (1985) found a good correlation between volumetric strain records and radon records in the Tokai district, west of the Izu peninsula and in the Izu peninsula. By comparing the radon record at SKE with the strain record at NGT for December 1982, Wakita et al. (1985) obtained an inversely
- 6

proportional constant 40 cpm / 1 x 10 between the two records. By assuming that the precursory change in radon concentration prior to the Izu-Oshima earthquake (Fig. 2-a) is caused by a

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change in volumetric strain, they calculated the amount of strain change from the change in radon content using the proportional
- 6

constant. Since the obtained strain change of 10 agrees with the observed strain change at NGT, Wakita et al. (1985) suggested
- 6

that simultaneous strain changes of the order of 10 took place throughout the Izu peninsula prior to the Izu-Oshima earthquake.

According to the above discussion, the Izu peninsula became -6
contracted by the order of 10 in early December 1977, and then in early January 1978, it was expanded and returned to the former state. This seems to be contradictory to the suggested drop of water pressure at well 1-4, and the lowering of water level at well 1-6 in December 1977. Also at Funabara, the lowering of water level was observed in December 1977 (Fig. 2-e), which will be discussed later.

A record of volumetric strain change at this station (NGT)
for a decade is shown in Fig. 7 together with records at stations FUJ and AJI (Earthq. Predict. Information Div., JMA, 1986). No reverse box-car type change other than that shown in Fig. 2-d is found in the NGT record for a decade. However, a few episodes of contraction with a similar amplitude were observed in 1978 without any subsequent large earthquake. Furthermore, similar but more distinct changes of reverse box-car type were observed at AJI in 1977, and at FUJ in 1978, but they were not followed by a large event. A total of eight dilatometers including those at the three stations mentioned above, were installed in 1975 and 1976. At other five staions no similar change of reverse box-car type was observed.

Sacks and Linde (1987) suggested that a tectonic event occurred near the Izu peninsula during 1978-1980 on the basis of observed borehole volumetric strain data. According to them, the largest changes due to this slow event were recorded at AJI: large compressional changes for more than one year starting around the end of 1977, followed by dilational changes, also for just over a year ending around June 1980. It was suggested that the start and end of this strain event are precursors of the 1978 and 1980 Izu earthquakes, respectively. Hagiwara(1985) observed similar temporal changes in gravity near AJI. However, similar changes were detected neither by repeated levelling surveys nor tide-gauge observations in the Izu peninsula. It is most likely that the observed gravity and strain changes are rather local changes.

During these changes, short-period changes consisting of a

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sudden expansion and subsequent slow contraction of exponential-decay type were repeatedly observed at AJI. It may suggest expansion of dilatometer due to inflow of hot water into the aquifer near the borehole. The observed long-term changes in strain and gravity might be also due to some unknown change in the local aquifer. According to Suyehiro(1985), a change in temperature by 0.1° C corresponds to a change in volumetric strain by 5 microstrain. Thus the observed strain changes of the order of 10 microstrain can be caused by changes in temperature by 0.2°C.

Water level anomalies at Funabara and Omaezaki

Reported intermediate-term precursors common to both the
1978 and 1980 events are the localized crustal uplift of the eastern coast, earthquake swarms, and the level of well water at Funabara (point e in Fig. 1) in the middle part of the peninsula. For the level of well water at Funabara, long-term data are unavailable and it is difficult to judge how significant is the reported anomaly shown in Fig. 2-e. However, it should be noted an amplitude of anomaly reported prior to the 1980 event is an order of magnitude smaller than that prior to the 1978 event (Fig. 8). The water level descended for a month in July, 1977 at a rate similar to that observed prior to the 1978 earthquake. Yamaguchi and Odaka (1978) attributed this descent to the artificial effect; hot water was used at a well 300m from the observation site for a swimming pool during summer holidays.

Reported water level anomaly at Omaezaki (point f in Fig. 1;
see Fig. 2-f) is also difficult to assess. Wakita et al.(1980b) claims that precursory lowering of the water level was also observed prior to the M=7.4 Miyagi-Oki earthquake of June 12, 1978 which took place 530km away from Omaezaki. The changes are superposed on a fast recovery of water level after pumping. The available record is two-year-long record contaminated with occasional changes of water level due to artificial pumping of water.

Crustal uplift

From the tide-gauge observations, the anomalous crustal uplift prior to the 1978 event was inferred to start at the beginning of 1975 (Crustal Dynamics Div., GSI, 1976; see Fig.11). The swarm activity started around August of the same year (Tsumura et al., 1977). Although the most seismically active area was not located at the center of the uplifted area (the area numbered 2 in Fig. 1) but on its southern flank and later moved to northeast of the area, swarm activity appears to be synchronized with an episodic uplift (Fig. 3).

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Fig. 3 includes the activity in 1930s when a localized uplift took place in approximately the same area with that started from the beginning of 1975. In spite of the general similarities among the three episodes shown in Fig.3, the precise location, dimensions, and lead time of uplift before the following event were different from event to event and there is no clear quantitative empirical correlation between the lead time or diameter of uplift and the earthquake magnitude. Table 1 summarizes earthquake sizes and the parameters of uplift episodes preceding each event.

Fig. 1 shows areas of uplift (areas where uplift exceeds one-half that of its maximum) #1, 2, and 3, prior to the 1930, 1978, and 1980 events, respectively, though an estimate of the uplift area prior to the 1930 earthquake is quite uncertain because of the data limited on one levelling line. Figure 9 shows vertical displacements during a period from 1967 to February, 1978 together with elevation changes along the east coast of the Izu peninsula during a period from February, 1973 to February, 1978. Several levelling loops in the peninsula and four tide gauge stations give a good control of the levelling data. Al-though coseismic effects of the 1974 and 1978 events are included in the data in the southern part of the peninsula, the figure shows that the maximum uplift attains more than 16 cm. As will be shown later in Fig. 11, this uplift mainly took place in one year, i.e., in 1975. Along the same levelling line on the east coast of the peninsula, three episodes of uplift are compared in Fig. 10. The size of the 1930 earthquake is the largest, but the extent of the uplift area appears to be the smallest.

By taking a difference of tidal heights between Ito tide gauge station located in the uplift area (Fig. 9) and Aburatsubo station 50km north east of Ito, some details of the history of uplift episodes can be seen (Fig. 11). Unfortunately we cannot get the details of the first uplift episode, because only the data of the repeated levelling surveys are available. There are several interesting features in Fig. 11. 1) The bulge did not collapse at the time of earthquake occurrence, suggesting some plastic deformation. 2) The large earthquakes did not occur when the uplift rate is the highest. Also the large earthquakes did not occur in the center of the uplifted area (Fig. 1). 3) There is no correlation between the precursor time and the earthquake magnitude. (Also there is no correlation between the spatial extent of the uplift and earthquake magnitude as was pointed out above.) These facts may suggest that the observed uplift episodes are rather indirectly related to the large earthquakes. It is most likely that each of episodic uplift and a large earthquake is a result of a common but unknown tectonic process.

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In spite of the fact that the spatial extent of uplifted area or the lead time of uplift is unrelated to the size of following earthquake, the lead time appears to be roughly correlated with the spatial extent (Table 1). By combining the spatial pattern of uplift (e.g. Fig. 9) with the time history of uplift (Fig. 11), we can estimate the maximum rate of uplift. The result listed in Table 1, indicates that uplift took place at almost the same rate. This may suggest that the three uplift episodes are resulted from the same physical process.

In its early stage, uplift was localized (see Fig. 9) and a pressure source at a depth of 10 km was suggested by Hagiwara (1977). As is shown in Fig. 1, a migration of uplift area can be seen from the 2nd (mainly in 1975 as shown in Fig. 3) to the third (mainly in 1978-1979) episodes. Furthermore the fourth (in 1982) and the fifth (in 1985-1986) episodes can be recognized from the tidal difference between Ito and other tide-gauge stations. These episodes of crustal uplift were associated with earthquake swarms. Fig. 12 shows the total vertical displacements duing the two latest episodes, a period from June-December, 1980 to May-July, 1986. A comparison of this figure with Fig. 1 indicates that the uplifted area further migrated to south. For a spherical pressure source, the depth of the source is equal to the surface distance from the center of uplift to the point where 35% of the maximum uplift is observed (e.g. Hagiwara, 1977). The .concentric pattern of vertical uplift in Fig. 12 also suggests a pressure source at a depth of 10km.

Superimposed on the concentric pattern of uplift in Fig. 12,
a pattern of broad uplift of the eastern coast and almost entire subsidence of the western coast can be recognized. Nakamura (1979) suggested that the present activity in the Izu peninsula is caused by a tectonic movement related to a bending of the peninsula on the Philippine Sea plate prior to its subduction at the Suruga trough. The broad pattern of deformation superimposed on the concentric pattern of uplift in Fig. 12 appears to be indicative of the bending.

Earthquake swarms and large earthquakes

Generally we observe a good time correlation between crustal uplift episodes and earthquake swarm activity. Since the third episodes of crustal uplift, which mainly took place around the end of 1978, short bursts of spatially-clustered seismic activity that lasted a few hours repeatedly took place several kilometers east to southeast of station AJI, which is shown in Fig. 1. Earthquakes in each burst-like activity are concentrated in a tight cluster about a few kilometer in diameter (e.g.

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Matsu'ura, 1983).

The mechanism of swarm earthquakes (e.g. Imoto, 1985; Imoto
et al., 1981) is generally consistent with the tectonic stress regime in the peninsula shown in Fig 11 (Nakamura et al., 1983). The burst-like nature of swarm activity might suggest repeated dike injection into the shallow part, but supportive evidence is unavailable in the focal mechanisms for that the dike injection is a direct cause of earthquakes. The earthquake mechanism consistent with the tectonic stress favors an idea that earth-quake swarms are caused by a rapid increase of pore pressure due to fluid (possibly gas) injection derived from magma body. The intrusion of magma into the shallow crust may also explain the spherical source of pressure assumed in the previous section to account for the concentric pattern of crustal uplift.

The mechanism of large earthquakes are also consistent with
the tectonic regime in this area (Figs. 1 and 13). This may suggest the bending of the Philippine Sea plate plays an important role in the present activity in the Izu peninsula. Shallow bending earthquakes are usually normal faults r but here we observe that the strike-slip mechanism is dominant in this region. Nakamura et al. (1984) interprete the difference as due to the collision (Matsuda, 1978) between the Philippine Sea plate and the plate bearing the Japanese islands, which is taking place in the immediate north of the peninsula. Lateral constraints appears to be greater than the usual bending stress due to the nearby collision so that the horizontal axis becomes the P axis rather than the intermediate B axis.

Magma intrusion and bending

Similarities between the present activity in the Izu peninsula, Japan and that in the Mammoth Lakes-Long Valley region, in Califrnia (e.g. Hill et al., 1985) should be noted. The areal extent and the amplitude of inflation of the resurgent dome in the Long Valley caldera is comparable to those of crustal uplift in the Izu peninsula. The Mammoth Lakes earthquakes of May 1980 and the Chalfant Valley earthquake of July 21, 1986 represent counterparts of the large Izu earthquakes, though the magnitudes of the former earthquake are smaller than those of the latter events. The temporal and spatial distributions of swarm earthquakes are quite similar to each other.

Major difference between the two regions lies in their tectonic settings. Also no migration of uplifted area was detected in the Long Valley region; uplift is mainly confined within the caldera. Shallow dike injection is suggested to accompany earthquake swarms in the Long Valley region while the

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secondary effect of dike injection is suggested for earthquake swarms in the Izu peninsula.

In the both regions, the major feature of the crustal uplift
can be explained by intrusion or inflation of magma. In the Izu peninsula, we also recognize that the bending of the plate plays an important role. It is most likely that each of a large earth-quake, episodic uplift, and an earthquake swarm in the Izu peninsula is resulted from some common but unknown tectonic process, which causes intrusion of magma and dike injection. One speculative idea may be that the bending of the plate is the ultimate cause and that the bending squeezes magma body from the lower part of the crust. The 1986 eruption of Oshima volcano lying 30km east of the Izu peninsula may also be related to this process.

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Table 1.

Main Shock M
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s 26
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dyne-cm years

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Half** Maximum
Width Uplift Rate
km cm/ month

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21
16

>1.6
1.3
1.7

26 Nov 1930
14 Jan 1978
29 Jun 1980

*The refereces are Abe(1978), Shimazaki and Somerville(1979), and Shimazaki(1980).
**Radius of an area where uplift exceeds one-half of its maximum.

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Tanna

Fault

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(^-1.5 month)
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SURUGA BAY

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i960
M=6.7

Izu-Oshima

I978
M = 7.0

Earthquake swarm

Volumetric strain

(^-1.5 month) o

Water level

(~1 month)

30km

Figure 1. Locations of intermediate-term anomalies discussed in this paper, uplifted areas, and major seismic activity in and around the Izu peninsula. The Tetters a to f attached to the closed diamonds show the locations of intermediate-term anomalies a to f shown in Figure 2. The shaded area marked g roughly shows an epicentral area of swarm earthquakes taking place in November 1977 (see Fig. 2-g). The lead times are given in the parentheses. The epicenters of major earthquakes are shown by the solid circles. The dotted, solid, and dot-dash-line curves numbered 1, 2, and 3, respectively, show uplifted areas prior to the large Izu earthquakes indicated by the corresponding numbers. The thin lines and the attached arrows indicate the earthquake faults and faulting motions of the large Izu earthquakes. The closed square indicates the location of dilatometer station AJI, whose volumetric strain record is shown in Fig. 7.

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Izu-Oshlma-kinkal earthquake
14 January 1976 (M7.0)

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SEP. OCT. NOV. DEC. JAN. FEB.
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' " 1978

Figure 2. Reported intermediate-term precursory changes prior to the Izu-Oshima earthquake of January 14, 1978 (Wakita, 1981; Wakita et al, 1985). Locations of these anomalies are shown in Figure 1 by the corresponding alphabets, a) Radon concentration changes at SKE (Wakita, 1981). b) Water temperature changes at 1-4 (Nagai et al., 1979). c) Water level changes at 1-6 (Kishi, 1979). d) Volumetric strain changes at NGT (Yamagishi et al., 1978). e) Water level changes after correcting for barometric pressure effect at Funabara (Yamaguchi and Odaka, 1978). f) Water level changes of a 500m well at Omaezaki at an epicentral distance of 90km (Wakita et al., 1980b). g) Seismic activity of the eastern part of the Izu peninsula. The number of earth-quakes per day observed at Kamata by the Japan Meteorological Agency.

42

CRUSTAL

UPLIFT

LARGE

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FORE-

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Figure 3. Three uplift episodes followed by large earthquakes. The numbers attached to the uplift episodes and the events correspond to the numbers shown in Figure 1. Fore-swarm and other earthquake swarms are also shown.

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Figure 4. Eight-year record of radon concentration of ground water at SKE (point a in Fig.l; taken from Fig. 6 of Wakita et al. (1986) and slightly modified). The arrows indicate the occurrence times of the M=7.0 Izu-Oshima earthquake of 1978, and of the M=6.7 East Izu (or Izu-Toho-Oki) earthquake of 1980. The attached numbers show epicentral distances of station SKE for these events. Please note that no significant change appear prior to the 1980 event which took place nearer to the station than the 1978 event.

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76 77 78 JASONDJFMAMJJA.S.ON.D.JFM

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(a) WATER TEMPERATURE AT 1-4

5m

Figure 5. (a) Record of water temperature at 1-4 (point b in Fig. 1; taken from Fig. 2 of Nagai et al.(1979)) and (b) schematic diagram showing configuration of the top part of the well and measuring point (taken from Fig. 24 of Nagai et al.(1979)). The water temperature was measured at point A after opening cock B for measurement. After the Izu-Oshima earthquake, cock B was kept open. Cock C was always open. Untill the end of November 1977, the water head was higher than 5m above the ground and hot water was pooled in the tank shown in the figure.

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Figure 6. Record of water level and precipitaion at 1-6 (taken from Fig. 2 of Kishi(1979)). Five-day average values of water level and total precipitaion in five days are shown. The float for measurement was completely stuck to the well wall during the period from April to the beginning of October 1977. The stability of measured water level during this period is not real but apparent. The effect of precipitaion lasting for less than a few days on the water level cannot be recognized (Kishi, 1979). The arrow indicates the occurrence time of the Izu-Oshima earthquake of 1978.

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- IOL WATER LEVEL AT FUNABARA(corrected for barometric pressure)

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WATER LEVEL AT FUNABARA

(corrected for barometric pressure)

1 APR 10 20 1 M AY 10 20
1980

20 30

Figure 8. Water level changes after correcting for barometric pressure effect at Funabara (Yamaguchi and Odaka, 1978; Yamaguchi, 1980).

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upheaval

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FEB 1978

Figure 9. Vertical displacements in a unit of cm during a period from 1967 to February, 1978 and elevation changes along the eastern coast. Ito and other three tide-gauge stations are shown by the closed squares. The levelling routes are also shown. The pattern of crustal uplift in the southern part of the peninsula is contaminted with co-seismic deformations associated with the 1974 and 1978 earthquakes.

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Figure 10. Elevation changes along a levelling line on the eastern coast of the Izu peninsula during the three episodes of crustal uplift (see Fig. 9 for the location of the levelling line).

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Ito-Aburatsubo

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1978 1979 1980

Figure 11. Uplift histories for the three episodes o± crustai uplift. For the first episode, changes in elevation difference between the levelling point 9328 (for its location, see Fig.
9) and point 9337 where the maximum uplift was observed (see Fig. 10.), are shown. For the second and third episodes, differences of tidal heights between Ito tide-gauge station located in the uplifted area (Fig. 9) and Aburatsubo station 50km northeast of Ito are plotted

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Figure 12. Vertical displacements during a period from June-December, 1980 to May-July, 1986. This period includes the fourth and fifth episodes of crustal uplift.

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Figure 13. Concentric trajectories of the T axes of the focal mechanism solutions of shallow earthquakes in the northern tip of the Philippine Sea plate (Nakamura et al., 1984). The radial broken and solid lines show P axis trajectories and general trend of the troughs, respectively. The open arrow indicates the direction of convergence between the Philippine Sea and Eurasian plates.

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Intermediate-term, Pre-earthquake Phenomena in California, 1975-1986, and Preliminary Forecast of Seismicity for the Next Decade

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Abstract

Intermediate-term observations preceding earthquakes of magnitude 5.7 or greater in Cali-fornia from 1975 through 1986 suggest that: 1) The sudden appearance of earthquakes in a previously inactive area indicates an increased likelihood of a significant earthquake in that area for a period from days to years; 2) Significant earthquakes tend to occur towards the ends of creeping fault segments ; 3) One large earthquake in a region increases the like-lihood of a subsequent significant event in the adjacent area; and 4) Marginal evidence for the occurrence of a regional deformation event suggests that such events increase the probability of earthquake occurrence throughout the entire area. A common element in many of these observed patterns appears to be the transmission and amplification of tectonic stress changes by the mechanism of fault creep, and suggests that surface fault creep is a sensitive indicator of changes in stress. The preceding criteria are used to construct a preliminary "forecast" of the likely locations of significant earthquakes over the next decade.

Introduction

A considerable body of observations has been accumulated over the last decade relevant to the intermediate-term prediction of earthquakes in California up to about magnitude 6.5. Possible intermediate-term precursory phenomena noted to date include variations in surfi-cial creep rates, in spatial and temporal seismicity patterns, and fluctuations in the rate of regional crustal deformation as measured by geodetic techniques. This paper reviews ob-servations preceding and following significant earthquakes in California, 1975-1986, seeking to determine whether these events were preceded by intermediate-term phenomena that could be recognized and used as a basis for qualitative or probabilistic forecasts of fu-ture earthquake occurrence. Based on the body of observations, and the emerging physical and geologic understanding of the observed phenomena, a set of qualitative, intermediate-term "forecast rules" is proposed for identifying likely locations for earthquakes magnitude 5.7 and above during the next decade, and for interpreting potential intermediate-term phe-nomena that may occur.

Table 1 lists all earthquakes in California with magnitudes greater that about 5.7 since 1975, and summarize the intermediate-term observations that preceded each. These obser-vations are discussed in detail below. Consistent with the pattern pointed out by Alien (1981), only one of these earthquakes is located on the San Andreas fault proper: the 1986 North Palm Springs earthquake (Figure 1). In contrast, by far the most active area of Cali-fornia during the last decade has been the Mammoth Lakes-Long Valley caldera region of eastern California.

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Observations

Oroville

The Oroville earthquake of August 1,1975 (Ms 5.7) occurred in the Sierra Nevada foothills of California, and was associated with displacement on the Cleveland Hills fault zone (Lahr et al., 1976; Clark et al., 1976). Although the earthquake was distant from any instrumentation designed for earthquake prediction studies, a local seismograph network was in operation because of the nearby dam and reservoir. Although there has been con-siderable discussion about whether the earthquake sequence was induced by the impound-ment of the reservoir, there can be little doubt that the earthquake was preceded by an intensification of seismic activity. From 1964 through June 27, 1975, only two earth-quakes were located within ten kilometers of Lake Oroville (Morrison et al., 1976). A foreshock sequence began on June 28, 1975 that included 25 locatable earthquakes, including one of magnitude 3.5 on June 28, and four foreshocks with magnitudes greater than 3.0 on August 1 (Morrison et al., 1976; Bufe, 1976; Lahr etal., 1976).

Santa Barbara

The Santa Barbara earthquake of August 13, 1978 (ML 5.1 Pas, ML 5.7 Brk) was located at a depth of 13 km and was caused by thrusting on a nearly horizontal, northward dipping, plane striking N80°W (Corbett and Johnson, 1982). The main shock was followed by 373 locatable aftershocks between August 13 and September 20. No surface fault rupture was reported. The Santa Barbara area is characterized by a rather high level of seismicity, with earthquakes commonly occurring in swarms (Corbett and Johnson, 1982). Indeed, the earthquake was preceded by a swarm about 25 km to the southeast, beginning in March 1978, which continued sporadically through July, and included a possible foreshock 4 hr before the August 13 mainshock (Whitcomb et al., 1979; Corbett and Johnson, 1982). Of possible interest from the view of intermediate-term observations is the apparent tendency for larger earthquakes to migrate westward in the western Transverse Ranges during the 1970's (Lytle Creek, ML 5.4, 1970; San Fernando, ML 6.4, 1971; Point Mugu, ML 6.0, 1973; Santa Barbara, 1978).

Coyote Lake

The August 6, 1979 Coyote Lake earthquake (ML 5.9) occurred within the Calaveras fault zone 110 km southeast of San Francisco (Lee et al., 1979; Uhrhammer, 1980). The rup-ture initiated at a right-stepping offset in the fault trace and propagated southeastward (Bakun et al., 1980). Detailed studies of the aftershocks by Reasenberg and Ellsworth (1982) indicate that the earthquake was associated primarily with slip on two sub-parallel, en echelon, near-vertical surfaces. Surface faulting associated with the earthquake occurred principally as discontinuous zones of cracking, across which displacement increased with

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time (Coppersmith et al., 1979; Herd et al., 1979; Rogers et al., 1982). King et al. (1981) interpreted trilateration measurements made in May and September, 1979, to indicate a co-seismic displacement of about 33 cm on a 21 km long rupture surface extending from 4 to 12 km depth. Lisowski and King (unpublished data) interpreted short-range trilateration network observatons to indicate a near-surface slip rate of about 13 mm/yr on the Calaveras fault prior to the earthquake, a coseismic slip of about 10 mm, and a post-seismic surface slip, through November 1980, of 40 to 100 mm. These observations suggest that the principal displacement at the time of the earthquake was confined to a depth greater than the baseline sampled by the short-range trilateration measurements.

Raleigh et al. (1979) noted that creep at Shore Road, along the Calaveras fault at the south-ern end of the aftershock zone, behaved anomalously for about three years prior to the earthquake (Figure 2). From 1971 to 1975, the fault crept episodically at an average rate of about 1 cm/yr. However, in 1976 creep seemed to stop, and virtually no displacement was observed until the time of the Coyote Lake earthquake. This has been interpreted as an example of "creep retardation" (Burford, 1976; Burford et al., this volume). Wyss and Habermann (this volume) also report a possible precursory seismic quiesence during this period, beginning about 22 months prior to the mainshock. It should be noted that other creepmeters along the Calaveras fault south of Shore Road recorded "normal" creep throughout this period (Schulz et al., 1982).

Imperial Valley

The Imperial Valley earthquake of October 15, 1979, is the largest earthquake (ML 6.6) to have occurred onshore in California since the advent of detailed and systematic geophysical observations for earthquake prediction. Although the observations may not be as com-plete or extend back in time as far prior to the earthquake as one might like, the earth-quake and the regional observations preceding and following it offer considerable interest. Detailed observations of seismicity with a local seismograph network began in 1973 (Hill et al., 1975), and served as the basis for detailed studies of the pre- and postmainshock seis-micity (Johnson, 1979; Johnson and Hill, 1982; Johnson and Hutton, 1982). These data have been augmented by master event relocations of earthquakes recorded on the regional network operating since 1932, and by earlier intensity observations, to construct a record of seismicity covering the period 1902-1984 (Doser and Kanamori, 1986). Earthquakes in the Imperial Valley occurred primarily on the Imperial, Brawley, and Superstition Hills faults. The largest of these was the 1940 earthquake (Ms 7.1), which, like the 1979 event, was associated with tectonic ground rupture on the Imperial fault. Although long baseline geodetic measurements in the Imperial Valley are complicated by the extreme flatness and broad expanse of the valley floor, measurements of the adjacent Salton network are available (Savage et al., 1981; Savage, 1983), as well as measurements of fault creep along the Imperial, Brawley and Superstition Hills faults (Brune and Alien, 1967; Goulty et al., 1978; Cohn et al., 1982; Louie et al., 1985).

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From of the view of intermediate-term earthquake prediction, several fascinating aspects of the observations in the Imperial Valley emerge. Earthquake swarms occurred in the El Centra area 2 to 5 years before both the 1940 and 1979 earthquakes (Figure 3) (Doser and Kanamori, 1985). Johnson (1979) noted the increase in seismicity that began in 1975, and commented on its similarity with intermittent earthquake swarms preceding the 1940 earth-quake. For about 3 months immediately prior to the 1979 earthquake, however, the level of seismicity in the area was remarkably low (Johnson and Hutton, 1982). This pattern of high seismicity followed by a sudden quiescence is considered significant relative to the behavior of background seismicity and may be an intermediate-term precursor to both the 1940 and 1979 earthquakes. In terms of spatial patterns, both events occurred along a sec-tion of the Imperial fault where the long-term secular creep rate diminishes from about 5 mm/yr to about 1 mm/yr (Louie et al., 1985).

Livermore

The Livermore earthquakes of January-February, 1980, occurred along the Greenville fault. The sequence began with a small foreshock (ML 2.7) about one and a half minutes before the main event (ML 5.9) on January 24 (Cockerham et al., 1980). Fiftynine after-shocks (ML>2.5) occurred over the subsequent 6 days. Several centimeters of right-lateral offset occurred along the Greenville fault for a distance of 6 km. Displacements increased in several places after the largest aftershock (ML 5.3) on 27 January (Bolt et al., 1981). Earthquake activity continued to migrate southward, with a small felt earthquake (ML 3.6) occurring on February 21,1980, about 10 km southeast of the second principal earthquake (near the intersection of the Greenville and Los Positas faults), and another felt event (ML 4.1) on August 24,1980, about 35 km south of the first principal earthquake, and near the southern boundary of Livermore valley (Followill and Mills, 1982).

Mammoth Lakes-Long Valley

The Mammoth Lakes earthquake sequence began on October 4,1978, with the occurrence of a magnitude ML 5.8 earthquake near Wheeler Crest, about 12 km south-southeast of Lake Crowley. Prior to the occurrence of this event, the interior of the Long Valley caldera area had been notable for its low level of microearthquake activity (Pitt and Steeples, 1975; Steeples and Pitt, 1976). Following a lull in seismicity that extended from October, 1978 to July, 1979, earthquake activity resumed, migrating northwestward towards the Long Valley caldera. This migration culminated in a series of four ML>6.0 earthquakes within 48 hrs on May 25 and 27, 1980 (Ryall and Ryall, 1981; Cockerham and Pitt, 1984). Sub-sequently, it was discovered that between July 1979 and September, 1980, the resurgent dome within the caldera had bulged upward about 0.25 m, and expanded outward, sug-gesting the inflation of a magma chamber at depth (Savage and Clark, 1982). After the May, 1980, sequence, intense swarms of earthquakes within the caldera continued, and an additional 9 cm of uplift of the resurgent dome was measured in 1982 and 1983 (Savage and Cockerham, 1984). Additional moderate size earthquakes have occurred, apparently

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associated with an ongoing process of magma migration. Persistence of the earthquake ac-tivity, and continued uplift within the caldera, suggest that the recent activity is part of a long-term process involving the redistribution of stress, and consequently, the likely occurrence of additional significant earthquakes in the coming years.

Victoria (Mexico)

The Victoria, Mexico, earthquake (Ms 6.4) occurred June 9, 1980, on the Cerro Prieto fault. The event is mentioned in this context because its rupture area is adjacent to that of the Imperial Valley earthquake, and seems to represent a southward extension of the fault-ing process associated with the Imperial Valley event, albeit on an en echelon fault (Wong and Frez, 1984; Silver and Masuda, 1985) (Figure 4).

Eureka

Although one of the largest earthquakes to occur in the northern California region since the great 1906 earthquake, little is known about the November 8, 1980 Eureka earthquake (Ms 7.2) because of its location offshore on a fault system accommodating the internal de-formation of the subducting Gorda plate. Aftershocks delineated a zone nearly 140 km long, trending N50°E, consistent with a nearly-vertical nodal plane of the mainshock focal mechanism indicating left-lateral motion on a northeast-oriented fault (Eaton, 1981). This style of deformation is similar to previous moderate to large size earthquakes that occurred in 1975 (ML 5.2) south of Eureka, and in 1976 (Ms 6.8) further offshore in the center of the Gorda plate (Smith et al., 1981).

Westmorland

The April 26, 1981 Westmorland earthquake (ML 5.7) was part of a swarm of more than 2,000 earthquakes that began in late April, 1981, and extended into early May (Hutton and Johnson, 1981). These events were located south of the Salton Sea, about 10 km north of the largest aftershock (ML 5.8, October 15, 1979) of the 1979 Imperial valley earthquake. Two subparallel planes oriented nearly perpendicular to the overall trend of the Brawley seismic zone were active, with each fault segment exhibiting its own cluster of foreshocks, a mainshock characterized by bilateral rupture and aftershocks defining a nearly vertical, northeast oriented fault plane. Focal mechanisms of the two largest events (magnitudes 4.1 and 5.7) were consistent with predominantly left-slip motion on the northeast-striking nodal plane. In addition, minor amounts of triggered right-slip surface displacement was found on both the Superstition Hills fault and the Imperial fault following the swarm sequence (Sharp et al., 1986). The sequence was located within the Brawley seismic area, adjacent to the northern end of the aftershock zone of the Imperial Valley earthquake (Johnson and Hutton, 1982), and north of the creeping section of the Imperial fault (Louie et al., 1985)(Figure 5).

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Codinga

The Coalinga earthquake of May 2, 1983 (ML 6.7) occurred about 33 km ENE of the San Andreas fault in an area dominated by active folds. Large numbers of aftershocks were generated, with at least seven larger than magnitude 5.0 (Eaton et al., 1983). Seismic evi-dence indicated nearly pure dip-slip motion on a low-angle (32°) thrust fault dipping to the southwest. Most of the aftershocks occurred at or above the fault plane and were bounded to the west by a secondary conjugate fault dipping steeply to the east. This conjugate plane was also apparently active during the earthquake sequence and was responsible for a num-ber of the largest aftershocks (including one as large as M 6.4) and generated the only observable surface fracturing (Hart and McJunkin, 1983). The Coalinga aftershock se-quence filled a previously quiescent zone defined by earlier aftershock clusters of moderate size earthquakes in 1975, 1976, 1980 and 1981 (Eaton et al., 1983)(Figure 6). Surface fault displacement occurred largely in association with aftershocks and as afterslip (Rymer et al., in press a and b). The Coalinga earthquake significantly affected creep on the San Andreas fault (Mavko et al., 1985).

Morgan Hill

The Morgan Hill earthquake (Ms 6.1) occurred April 24, 1984, along a 30-km section of the southern Calaveras fault extending northwestward from the inferred rupture zone of the Coyote Lake earthquake (Cockerham and Eaton, 1984; Bakun et al., 1984) (Figure 7). It was recognized in 1980 that an earthquake of about this size had occurred on this section of the fault in 1911, and that an intermediate-term seismic slip deficit along this section had accumulated (Bakun, 1980). No short-term prediction was made, however. Although no unambiguous surface fault displacement was observed at the time of the earthquake, the creepmeter at Shore Road, about 50 km southeast of the epicenter along the Calaveras fault, showed a 0.11 mm offset coincident with the earthquake, and a 12.9 mm creep event be-ginning about 23 minutes after the event (Schulz, 1984) (Figure 8). The Coyote Lake and Morgan Hill earthquakes are among the best examples of the interplay between seismic and aseismic slip (Bakun et al., 1986) and of the loading of adjacent parts of a fault by an initial failure. Wyss and Habermann (this volume) have identified a precursory seismic quies-cence beginning about 23 month prior to the mainshock.

Kettleman Hills

The ML 5.9 North Kettleman Hills earthquake of August 4, 1985 occurred at a depth of 12 km along the western edge of the Great Valley and about 17 km southeast of the epicenter of the Coalinga earthquake (see Figure 6). Aftershocks of the 1985 event abut the southeastern end of the Coalinga aftershock zone. The length of the 1985 aftershock zone was about 20 km, as compared to about 30 km for the Coalinga event. The mainshock focal mechanism is consistent with low-angle thrusting on a plane striking N52°W and dipping to the southwest at 12° (Eaton, 1986). This earthquake sequence appears to be a

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continuation of the southeastward migration of events through the hills that flank the western edge of the Great Valley which began in 1974.

North Palm Springs

The July 8, 1986 North Palm Springs (ML 5.9) earthquake occurred on the Banning strand of the southern San Andreas fault at the northern end of Coachella valley. The earthquake generated an unusually rich aftershock sequence with earthquakes as large as 4.7 occurring as late as October 14. The aftershocks define a zone about 15 km long and 10 km wide that dips to the north and east at about 50° (Jones et al., 1986). Rupture was apparently confined between depths of about 5 and 13 km, and propagated to the northwest and southeast, as well as up and down dip. Surface trace fracturing was localized to about a 10 km length of the Banning, as well as a short (about 1.5 km) length of the Mission Creek fault near the epicenter, and possibly another short section of the Garnet Hill fault at the southern end of Whitewater Canyon (Sharp et al., 1986). Triggered slip along the San Andreas fault as far south as the Mecca Hills followed about a day later (Williams et al., 1986). The earthquake occurred at the northern end of the creeping section of the southern San Andreas fault, which exhibits a creep rate of about 2 mm/yr along the Banning strand at Devers Hill (Louie et al., 1985).

The earthquake occurred at the eastern end of a zone of deep seismicity in San Gorgonio pass. Prior to the mainshock, the fault segment responsible for the earthquake appeared nearly aseismic, even at the microearthquake level. What few earthquakes there were in the immediate area of and prior to the mainshock occurred either as left-lateral events on secondary transverse faults in the hanging wall (upper plate) or at depths greater than about 10-12 km to the west (in the lower plate) (Nicholson et al., 1986a). In January of 1985, a swarm of magnitude 3.5-4.0 earthquakes occurred, outlining a northeast-trending feature approximately 8 km east of the Mission Creek fault (Figure 9)(Norris et al., 1986). This swarm occurred at about the southeastern end of the relocated 1948 Desert Hot Springs (ML 6.5) aftershock sequence (Nicholson et al., 1986b), and about 10 km from the southeastern end of the 1986 rupture zone (Figure 9). This activity, as well as the 1986 North Palm Springs earthquake, may be similar to the regional increase in seismicity that occurred in the 7-13 years prior to the 1948 earthquake (Richter et al., 1958; Sykes and Seeber, 1986), and may thus presage future moderate size activity in this area for the next several years.

Oceanside

The July 13th earthquake (Ms 5.8) occurred off Oceanside in the San Diego trough along a possible continuation of the Palos Verdes Hills fault. The earthquake apparently occurred along one of the few segments of the fault zone known to disturb late Quaternary sedi-ments, based on high-resolution acoustic profiles (Clark et al., 1986). More importantly, however, the earthquake was preceded by a burst of seismicity on 16 June, 1985. This ac-

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tivity was part of regional increase that included several felt earthquakes within the city limits of San Diego on the 17th (Figure 10) (Norris et al., 1986; Heaton, this volume). Seismic activity in the San Diego area continued at a relatively high rate until the 1986 event (Heaton, this volume).

ChalfantValley

The Chalfant Valley earthquakes of July 20 (ML 5.9), July 21 (ML 6.4), and July 31 (ML 5.8), 1986 occurred as part of a sequence of earthquakes in the northern part of Owens Valley beginning July 3,1986 (Cockerham and Corbett, in press). Epicenters of earthquakes in the sequence cover a "boomerang" shaped area about 24 km long and 8 km wide trending south in the northern part of the area and southsoutheast in the southern part, dipping about 60 degrees to the southwest. Preliminary studies of the sequence (Cockerham and Corbett, in press) indicate three aspects of this sequence of great interest from the point of view of intermediate-term earthquake prediction. First, the sequence oc-curred in the northern part of what had been identified as the White Mountain seismic gap (Hill et al., 1985). Second, the sequence is closely related in space (about 30-50 km to the east southeast) from the location of intense seismic activity in the Mammoth Lakes-Long Valley area (Cockerham and Corbett, in press) (Figure 11). Third, the largest earthquakes were preceded by a clear sequence of foreshocks (Figure 12), regarded at the time as un-usual and worthy of special field investigation (Cockerham and Corbett, in press). In ad-dition, both the North Palm Springs and Oceanside earthquakes occurred during the early part of this sequence.

The 1979 Strain Event

Savage et al. (1981) and Savage and Gu (1985) have presented and critically reviewed evi-dence for a regional strain event in southern California in 1979. The event was observed as an apparently anomalous contraction perpendicular to the average trend of the San Andreas fault system that accumulated in 1973-1978, and was abruptly released in 1979. This anomaly contrasts with the long-term strain accumulation as simple right-lateral shear at the rate of 0.2 jistrain/year parallel to the fault. Although most clearly observed at Palmdale, marginal evidence from five large-aperture trilateration networks (Savage et al., 1981) sug-gested that the event did indeed cover a large part of southern California, extending about 400 km northwestward from the Imperial Valley, moving northwestward with time. Sub-sequent measurements, especially those made in 1983-1984, show rather more scatter than observed previously, and thus have reduced confidence in the reality of the regional anomaly (Savage et al., 1986). However, Savage and Gu (1985), focussing on the data from the Palmdale area, conclude that, although evidence for the existence of the anomaly is not overwhelming, their analysis, taken together with the apparently anomalous changes near Palmdale in gravity (Jachens et al., 1983) and water level (Merifield and Lamar, 1981), suggests that the anomaly is real (Figure 13).

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Although Savage et al. (1986) were inclined to reject the apparent 1977-1978 anomaly dilatation in the observed in the Salton and Anza networks as not being significant (Figure 14), it is interesting to note that the alignment array at Dixieland, in the southwest corner of the Salton network, was also showing anomalous behavior during this period, with the line Supe Dixie showing a relatively large amount of shortening. The alignment array at Dixieland, which showed 47 mm of right lateral slip from 1970 to 1976, reversed and showed 11 mm of left-lateral slip during the interval 1976 to 1977(Figure 15).

These apparent strain anomalies are particularly important because they coincide in time with a rather remarkable change in the seismicity of southern California noted by Bufe and Toppozada (1981). From the beginning of 1973 through mid-1978, only one earthquake of magnitude (ML) 5.5 or greater occurred in southern California. In contrast, from mid-1978 through mid-1980, there were seven such earthquakes, excluding aftershocks. The largest of these earthquakes was the 1979 Imperial Valley earthquake (ML 6.6). Bufe and Toppozada (1981) point out that the complete absence of earthquakes of magnitude 5 or larger in all of onshore California for the three year period mid-1975 to mid-1978 is unique in the 85-year seismic history of California. Savage et al. (1981) suggested that the possi-ble strain anomalies observed on the trilateration networks may have indicated a regional perturbation of strain, associated with, if not responsible for, the subsequent occurrence of the earthquakes.

Discussion of Observations

The overall impression obtained from a review of these observations is that there are sys-tematic, if poorly understood, causal relationships among at least some of these earth-quakes, and possibly with fluctuations in the rate of regional strain. The apparent north-ward migration of earthquakes along the Calaveras fault, the apparent southward migration of seismicity along the western margin of the Great Valley (including the Coalinga and Kettleman Hills earthquakes), the earthquakes at both ends of the creeping section of the Imperial and Brawley fault zones, and the close spatial association of the intense seismicity in the Long Valley and Chalfant Valley areas, all suggest some process by which stress is transferred from one part of the region to an adjacent part. This interpretation is strength-ened by the observation that of the earthquakes along fault zones known to be creeping, most occurred near the ends of the creeping zones (Figure 16),where stress concentration is expected. These earthquakes include the Coyote Lake and Morgan Hill earthquakes at the northern end of the most rapidly creeping part of the Calaveras fault, the Imperial and Westmorland earthquakes at the south and north ends of the creeping parts of the Imperial and Brawley faults, and the North Palm Springs earthquake at the northwestern end of the creeping section of the southern San Andreas fault.

While these patterns may be hard to prove statistically, the proper question for analysis is not whether these seismicity patterns are statistically significant in the abstract, but rather, is there a physical model within which this behavior is understandable. The challenge of

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intermediate-term prediction seems to be to find the right model for the interpretation of these observations. The strength of fault zones and fault creep seem to be important elements of such a model.

Observations of fault creep are now being made directly with creepmeters and geodetic techniques, and indirectly with strain meters and water wells. An important characteristic of fault creep is that it seems to be a very sensitive indicator of changes in tectonic stress on both local and regional scales as indicated by creep steps, and by changes in the character and rate of fault creep induced by the occurrence of nearby and regional earthquakes. With the possible exception of the 1979 strain event discussed above, long baseline trilateration measurements at least with the density at which they are currently surveyed seem rela-tively insensitive to regional effects associated with earthquakes in this magnitude range, whether the effects are imagined to come before, during, or after the earthquakes. Simi-larly, the extremely sensitive and stable laser strainmeters at Pinon Flat have apparently not detected any regional strain fluctuations of a tectonic nature, except for coseismic offsets. Both the long-baseline trilateration and laser strain meter results not withstanding, the be-havior of fault creep (c.f. Sieh, 1982) seems to reflect changes in stress resulting from sig-nificant earthquakes in southern California at distances up to 90 km . Fault creep, there-fore, seems to be the most sensitive indication currently available of changes in tectonic
stress.

How can this be? Perhaps a useful approach is to consider, the earth's crust in California not as an elastic half space, but as a mosaic of relatively stiff elastic plates or blocks, sepa-rated by yielding, relatively weak, fault zones similar to the model discussed by Bilham and Beavan (1979). In this model, deformation is concentrated in the fault zones. Measure-ments made within the stiff plates reflect relatively little strain of a tectonic nature, and long-baseline trilateration measurements tend to average deformation across both plates and fault
zones.

Computational methods are now available to assess changes in stress associated with both creep and earthquakes, and to assess how these stresses are redistributed and concentrated. There has been a tendency in the past to dismiss the importance of the static stress change associated with an earthquake at distances of more than a few km, because of the rapid at-tenuation of the change with distance from the rupture. Typically the change in stress is less than a bar at a distance of 10 km. However, studies of sympathetic slip, triggered creep events, and changes in creep character, all suggest that the creep behavior of faults is responsive to static stress changes of this order or less, and imply the existence of quite weak material within creeping faults.

Another perplexing factor has been the physical basis for the delay between earthquakes on adjacent portions of a fault zone. As demonstrated by Rybicki et al. (1985), the dynamic stresses associated with the propagation of elastic waves away from an earthquake, are much larger than the static stress changes. The observations, however, indicate that failure

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is not immediately triggered by the propagation of elastic waves, but is delayed days to years. This observation, taken together with the evidence on the weakness of creeping faults, and results from the recently developed computational approaches, suggest the im-portance of the mechanism of "creep loading," or "stress amplification."

Consider an asperity within a creeping fault zone (Figure 17). At the time of an earthquake some distance away, the creeping portions of the fault will generally be close to a state of static equilibrium with the regional tectonic stress. That is, if the rheology of the creeping part of the fault zone is such that the creep rate is proportional to the applied stress, then at static equilibrium, the creep rate is zero because the applied stress is zero, since the tectonic stress is just balanced by the dislocation stress associated with the distribution of displace-ment on the fault. Segall and Harris (1986) give a clear picture of how an asperity of this type (namely the one at Parkfield) is loaded by secular creep. From an assumed geometry of the asperity/creeping fault system, and an assumed change in regional stress, the distri-bution of displacements on the creeping part of the fault zone can be calculated (Wesson, in preparation). In addition, the increase in stress on the asperity, owing to both the change in regional stress and subsequent stress redistribution by creep can be calculated. Assuming, as above, that the creep rate is directly related to the applied stress, creep will initiate on the creeping parts of the fault even for very small changes in stress. The stress change on the asperity following creep will be higher than the change in the regional stress. Roughly, if the proportion of the area of the asperity on the fault surface to the total area of the surface (that is, including both asperity and creeping parts) is A, then the change of stress on the

asperity at equilibrium, AT is given by asp
)ATreg

where AT is the change in the regional stress. reg

If the stress on the asperity as creep proceeds to equilibrium exceeds the strength of the asperity, a triggered earthquake (or possibly triggered creep) will ensue. In this model, the delay between an earthquake (or creep event), and subsequent triggered earthquakes (or creep events) is related to creep in the vicinity of the triggered event. If a constitutive law for this creep can be deduced or assumed, then the stress amplification can be simulated numerically (Wesson, in preparation).

Implications of Observations and Interpretation

Observations of earthquakes in California over the last decade, and the physical arguments discussed above motivate the following preliminary set of temporal and spatial "forecast rules" for assessing future seismicity based on that of the recent past:

A sudden increase in seismic activity in a previously quiet area must be taken
seriously as an intermediate-term precursor (Oroville, Mammoth, North Palm Springs, Oceanside, Chalfant Valley), and should be reflected as a probability

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gain in assessing the intermediate-term seismic potential of the surrounding area.

Larger earthquakes in California along faults exhibiting fault creep tend to oc-cur at the ends of creeping sections, or where the creep rate is markedly reduced (Coyote Lake, Imperial Valley, Morgan Hill, Westmorland, North Palm Springs).

Large earthquakes tend to occur adjacent to previous large earthquakes, repre-senting either an extension of the rupture (Morgan Hill), a migration, or perhaps as a response (commonly delayed by creep) to static stress change (Mammoth, Kettleman Hills, Chalfant Valley, Westmorland, Mount Lewis). Therefore, the occurrence of any significant earthquake increases the intermediate-term proba-bility of future earthquakes on adjacent fault segments and in the surrounding
area.

There is marginal evidence for periods of high regional deformation (the 1979 Strain Event, the period of intense seismic activity during 1979-1980, the period
of intense activity during July 1986). Therefore, any evidence for a period of
similar high regional deformation (a regional strain event or the occurrence of an earthquake with a magnitude above about 5.7) increases the intermediate-term probability of future earthquakes throughout the region.

A Preliminary, Ten-year Forecast

The characteristics of California seismicity as described above, bolstered by previous ob-servations of California earthquakes and the concept of creep loading as a physical basis for understanding patterns of earthquake activity, permits a crude attempt to forecast Cali-fornia seismicity for the next decade. It must be stated at the outset, however, that since not all earthquakes in the last decade fit the deduced "forecasting rules" owing to lack of information (for example, as yet undiscovered creeping faults) or to lack of understanding (the forecast rules are inadequate or simply wrong) any forecast based on those rules will likely miss several earthquakes. With that caveat, a preliminary, qualitative, ten-year forecast for earthquakes larger than magnitude about 5.7 in California, may be proposed. The strategy behind the forecast has two parts: first, the identification of the regions in California based upon the observations up to November, 1986 where the probability of an earthquake of magnitude 5.7 or greater has a significantly higher probability than the long-term average over the coming decade; and second, the use of temporal aspects of the "forecast rules" listed above, to update and modify the forecast as events unfold.

The spatial part of the forecast focuses on the initiation points of rupture, rather than on the anticipated rupture dimension or magnitude. However, rupture dimensions and anticipated magnitudes may, in many instances, be estimated by comparison with the long-term fore-casts based on fault segmentation arguments (e.g. Lindh, 1983; Sykes and Nishenko,

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1984). Intermediate-term analysis of spatial and temporal patterns of seismicity based on slip budget calculations (e.g. Wesson et al., 1973; Bakun et al., 1980; Bakun, 1981; Bakun et al., 1986) are useful in cases where faults are well mapped, earthquakes appear confined to the main fault, and where a record of seismicity, creep and geodetic observa-tions exist. However, as is clear from the last decade, a relatively small proportion of the earthquakes in California can be expected to occur in such circumstances over the next decade. It therefore seems useful to take a more generalized approach. The spatial part of the forecast (Figure 18) is based principally on the currently observed distribution of fault creep, and on the pattern of seismicity over the last decade.

The specific areas of high probability for earthquakes in the next decade are discussed in detail below. It must be emphasized that the forecast is intended to be modified according to the temporal rules as possible intermediate-term phenomena are observed. Further, there is a virtual certainty that significant earthquakes will occur in locations not identified as high probability on this list. Therefore, this list should not be construed as an argument to re-duce vigilance or earthquake preparations in areas not identified. The discussion is keyed to the letters indicated on Figure 18.

Areas A and B North and south ends of the Hayward fault

The Hayward fault is actively creeping along most of its length, apparently concentrating stress at both ends. No earthquake with tectonic ground breakage has occurred along the fault since 1868. The occurrence of the earthquakes during the last decade on the southern Calaveras fault, discussed above, suggest an additional concentration of stress in this re-gion.

Area C-Central Calaveras fault and area east of Calaveras fault south ofLivermore Valley

The pattern of earthquakes on the southern Calaveras fault (1979 Coyote Lake, 1984 Mor-gan Hill) suggested a northward migration and concentration of stress in this region. A straightforward interpretation would have been to suggest a future earthquake on the Calaveras fault north of the rupture zone of the Morgan Hill earthquake. However, the oc-currence of the 1986 Mount Lewis earthquake north of the Morgan Hill rupture zone, but east of the Calaveras fault indicates that while the interpretation of high stress was sound, the understanding of faults was simplistic. Further, there may be a gap between the rupture zones of the 1980 Livermore and 1986 Mount Lewis earthquakes. Therefore, the broad region extending from (and including) the Calaveras fault eastward into the gap between the Mount Lewis and Livermore earthquakes is identified. Note that this region is contiguous with area B, at the south end of the Hayward fault.

Area D--San Juan Bautista-Watsonville-Corralitos

Although very active in the 1960's and early 1970's with earthquakes in the magnitude 5

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range, this area at the northern end of the creeping section of the San Andreas fault was relatively quiet in the period 1975-1986. Earthquakes in the early 1980's along the section of the San Andreas fault south of San Juan Bautista, suggesting a repetition of the sequence in the early 1970's (Ellsworth, this volume), provide evidence of high stress in this region.

Area E-Parkfield-Cholame

The Parkfield area meets the qualitative criteria used in this paper for a forecast, because it is located at the southern end of the creeping section of the central San Andreas fault seg-ment, and did not have a significant earthquake in the previous ten-year period. Of course, the arguments for an earthquake at Parkfield in the next decade are much more extensive (c.f. Bakun and Lindh, 1985 a and b) than suggested by these simple "forecast rules" alone.

Area F-South ofKettleman Hills

The pattern of seismicity along the western margin of the Great Valley described by Eaton (1983,1986) (Figure 6) suggests a southward migration of significant earthquakes in this region, as well as the presence of clusters of small earthquakes south of Kettleman Hills. Consequently, this area may be regarded as one with a heightened probability for a signifi-cant earthquake.

Areas G~Transverse Ranges west of Santa Barbara

Geologic evidence and historical seismicity, along with rather tenuous evidence of west-ward migration of significant earthquakes in the western Transverse Ranges, suggest the possibility of a significant earthquake in this region in the next decade. The level of confi-dence associated with this forecast is rather low, however.

Areas H and I-West and east ends of the Garlock fault

The geologic evidence for recent displacement (Clark, 1973), the existence of 4 mm/year of fault creep along the western portion of the fault (Louie et al., 1985), the presence of low-level, but continuing seismicity, particularly at the eastern end of the creeping segment {Astiz and Alien, 1983), and the absence of a significant earthquake in historic time suggest that this fault could generate future earthquakes, and that significant stress has accumulated at the two ends of the creeping segment.

Area J-Eastern San Gorgonio Pass-Desert Hot Springs area

The occurrence of the 1986 North Palm Springs earthquake, and the observations of creep along San Andreas fault to the southeast, indicate that this region has a high potential for generating a moderate to large earthquake in the near future. The criterion of "adjacency" is

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particularly important in this region, because of the complicated geologic structure, and the precedent of the occurrence of several earthquakes in the magnitude 5-6 range in this area in the late 1940's (Sykes and Seeber, 1986). Earthquakes in this range should be anticipated during the next decade in the region surrounding the aftershock area of the North Palm Springs earthquake.

Area It-Eastern and southern margins ofSalton Sea

The creep extending northward along the San Andreas fault from Mecca Beach, and the creep and large earthquakes south of the Salton Sea during the last decade, suggest that this region has been loaded from both north and south. Unfortunately, the tectonic connection between the San Andreas fault along the eastern margin of the Salton Sea, and the Brawley and Imperial fault systems to the south, is poorly understood. Nonetheless, and despite the lack of large historic earthquakes, this general region, on the basis of the slip occurring in adjacent areas seems to be a likely candidate for a significant earthquake in the next decade.

Areas L and M-Southern San Jacinto fault zone

For the last century, the San Jacinto fault has been the most active segment of the San An-dreas fault system in southern California for earthquakes greater than magnitude 6.0. Al-though at least 80% of the fault zone has ruptured in the last 95 years, certain sections may still be capable of generating significant earthquakes in the near future. The area south of Anza, where the fault zone is comprised of a complicated series of fault splays, including the Coyote Creek, Buck Ridge, Clark and Horse Canyon faults, has high seismicity and experienced a moderate size (ML 5.5) earthquake in 1980. This segment of the fault zone also is at the northern end of the creeping part of the Coyote Creek fault (Louie et al., 1985). For these reasons, the area that includes Coyote Ridge, the southern end of the Anza seismic gap, and the many splays of the San Jacinto fault zone, is a likely candidate for a future moderate to large earthquake. In addition, because creep apparently diminishes towards the southern end of the San Jacinto fault zone (Louie et al., 1985), and because seismicity on the Brawley seismic zone and the Imperial fault has been high over the last decade, the Superstition Hills and the Superstition Mountain fault are also considered po-tentially areas for future seismicity. Lack of knowledge about creep in this area, and about how plate motions are transferred from the southern end of the San Jacinto fault zone to the faults in the Imperial Valley, reduces confidence in this forecast.

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Area N~Mammoth Lakes-Long Valley area

The tectonic and volcanic processes that gave rise to significant earthquakes in the last eight years give little sign of abating. The intensity and duration of this activity suggests a high probability of additional significant earthquakes in this region over the next decade. Al-though the 1986 Chalfant Valley earthquakes suggests migration of this activity to the southeast, possible activity to the north in Mono Valley should not be ruled out.

Area O~Northern Owens Valley-White Mountains region

The extension of the intense seismicity in the Mammoth Lakes-Long Valley area over the last decade to the east and south, as indicated by the 1986 Chalfant Valley sequence, sug-gest that additional significant earthquakes are possible in the regions around Chalfant Val-ley, part of the White Mountain seismic gap (Hill et al., 1985). Of particular concern may be the regions between the rupture associated with the Chalfant Valley earthquake and that of the 1872 Owens Valley earthquake about 25 km to the south, and between the Chalfant Valley rupture and that of the 1934 Excelsior Mountain earthquake about the same distance to the north.

Area P~Oceanside and San Diego

The increase in seismicity that began in June, 1985, and culminated in the 1986 Oceanside earthquake, indicates an increased potential for future moderate size earthquakes in the off-shore region of San Diego, adjacent to the 1986 Oceanside event (Heaton, this volume).

Additional areas of concern

The "forecast rules" outlined in this paper for identifying intermediate-term preseismic phe-nomena focused as they are on fault creep and recent adjacent earthquakes have not identified any areas of enhanced probability in the greater Los Angeles area. Experience, however, urges caution on the forecaster to avoid giving a false impression of relative safety in this region. Note, in particular, that the 1971 San Fernando earthquake, although it was preceded by some local seismicity (Wesson and Ellsworth, 1973), would not have been forecast by these criteria.
In addition, fault creep has been suggested along some faults north of San Francisco Bay (c.f. Wesson et al., 1975). If additional information were available, it seems likely that additional areas in that region would meet the criteria discussed in this paper.

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Approach to quantitative assessment

An important future step will be to translate the qualitative forecast described above into quantitative, probabilistic terms. Possible directions for such an analysis are provided by review of Tables 1 and 2, which tabulate the phenomena preceding past earthquakes and potentially precursory phenomena to future earthquakes. Qualitatively these tables also provide some assessment of the uncertainties in the forecast itself. Of the 28 earthquakes listed in Table 1,16 might be called "independent," that is eliminating clear aftershocks and counting the sequence at Mammoth only once. Four of the 16 "independent" earthquakes exhibited no clearly identifiable precursory phenomenon (Santa Barbara, 1978 Mammoth, Livermore and Eureka). Five (possibly six) of the 16 events occurred at the ends of creeping segments. Ten areas are identified in Table 2 on the basis of location at the end of a creeping fault segment. Since the locations of the ends of the creeping segments are not known to vary in time, areas selected for future earthquakes by this criterion alone would have been selected in previous time period. Indeed, for completeness the ends of the creeping segments of the Imperial and Calaveras faults should be added, for a total of fourteen possible locations over the long term. During the last decade five of the 16 "independent" earthquakes occurred in four of these locations.

Thus, if the experience of the last decade can be considered typical a dubious assump-tion then, based on this very small sample, the following quasi-statistical statements can be offered. About one-fifth of the "independent" earthquakes in the next decade will occur without recognizable intermediate-term precursory phenomena. About one-third of the "independent" earthquakes will occur at the end of creeping fault segments, and about one-third of the locations identified in Table 2, based on this characteristic alone, will be the lo-cations of significant earthquakes in the next decade. Thus the probability of occurrence in any of the forecast areas is about one-third. Or, in terms of expectation, only about one-third of the areas forecast in Table 2 should be expected to give rise to a earthquake in the next decade. Appropriate statistical analysis of all the intermediate-term phenomena de-scribed above, particularly adjacent seismicity, and considering seismicity extending over a longer time period, should enable quantitative statements with much improved levels of confidence.

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Conclusion

A review of the significant seismological and geophysical observations in California over the last decade suggests that recognizable, but poorly understood, phenomena can be identified prior to major California earthquakes and may thus provide grounds for future intermediate-term earthquake prediction. Understanding the processes by which stress is transferred and redistributed through the Earth's crust probably fault creep is needed to develop a physical foundation for the interpretation of these observations. A set of "forecast rules" is discussed, summarizing the intermediate-term, precursory, phenomena observed in California over the last decade. These rules are then applied to assess likely locations of potential large earthquakes in California over then next decade. Whether these rules are valid, will, in large part, be determined by the extent to which the forecasts prove
accurate.

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Earthquake

Oroville
Santa Barbara
Mammoth
Coyote Lake
Imperial Valley
Imperial Valley
Livermore
Mammoth
Mammoth
Mammoth
Mammoth
Mammoth
Mammoth
Victoria (Mexico)
Eureka
Westmorland
Mammoth
Coalinga
Coalinga
Morgan Hill

Mammoth
Kettleman Hills
Mount Lewis
North Palm Springs
Oceanside
Chalfant Valley
Chalfant Valley
Chalfant Valley

Date

750801
780813
781004
790806
791015
791016
800124
800525
800525
800525
800525
800526
800527
800609
801108
810426
810930
830502
830722
840424

841123
850604
860331
860708
860713
860720
860721
860731

Time

2020
2254
1642
1705
2316
658
1900
1633
1649
1944
2035
1857
1450
328
1027
1209
1153
2342
239
2115

1808
1201
1155
920
1347
1429
1442
722

Ms-

Latitude

39-26.15
34-21.06
37-30.83
37-06.59
32-36.82
33-00.82
37-49.58
37-35.38
37-38.68
37-32.03
37-36.60
37-32.40
37-28.25
32-11.12
41-07.03
33-05.91
37-35.21
36-13.96
36-14.44
37-18.81

37-27.70
36-09.93
37-28.98
33-59.90
33-00.00
37-33.92
37-32.30
37-28.28

USGS surfa<

Longitude

121-31.37
119-42.00
118-40.96
121-30.68
115-19.09
115-33.31
121-47.16
118-49.57
118-53.97
118-49.06
118-51.60
118-53.40
118-48.15
115-04.55
124-39.89
115-37.90
118-52.24
120-18.57
120-24.53
121-39.39

118-35.57
120-05.09
121-41.40
116-36.90
117-48.00
118-26.27
118-26.59
118-22.11

Depth
(km)

9
7

7
12
9
8
7
7
14

16
5
2
4
5
10
7
10

13
6
8
6

9
11
9

Magnitude

5.7 Ms
5.6 Ms.5.7 ML
5.1 Ms,5.8 ML
5.7 Ms,5.9 ML
6.9 Ms
5.7 Ms
5.9 Ms
6.1 Ms
6.0 ML
5.8 Ms
5.7 ML
5.7 ML
L 6.0 Ms
6.4 Ms
7.2 Ms
6.0 Ms
5.8 Ms
6.5 Ms
5.7 Ms
6.1 Ms

5.7 Ms
5.9 Ms
5.5 Ms.5.7 ML
6.0 Ms
5.8 Ms
5.6 Ms,5.9 ML
6.2 MS.6.4ML
5.2 Ms.5.8 ML

Seismicity
increase

X

X
X

X
X
X
X
X
X

X

X

X?
X
X
X
X

:e wave magnitude, ML- Berkeley local magnitude

Adjacent
seismicity

X?

X

X
X
X
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X
X
X

X
X
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X
X

X
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X

X
X
X

End of
creeping
fault

X
X
X

X?

X

X

Creep
retardation

X

Quiescence

X?

X?

Uplift

X
X
X
X
X
X

X

X

Strain
event

X?

Observations and comments

Seismicity increase
Westward migration of seismicity in Transverse Ranges?
?

End of creeping fault, creep retardation, quiescence? End of creeping fault, preceding swarms, strain event?
Aftershock
?

Seismicity increase, adjacent seismicity, uplift

"

Adjacent seismicity
?

End of creeping fault?, adjacent seismicitv Seismicity increase, adjacent seismicity, uplift
Low level seismicity?
Aftershock, adjacent seismicity
End of creeping fault, adjacent seismicity,
northward propagation of seismicitv, guiescence? Seismicity increase, adjacent seismicity, uplift Adjacent seismicity, southward migration of seismicity Adjacent seismicitv. northward migration of seismicitv Seismicitv increase?, end of creeping fault Seismicity increase?
Seismicity increase, adjacent seismicity

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Table 1

Area

A
B
C
D
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F
G

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P

Location

North end of Hayward fault South end of Hayward fault
Central Calaveras fault and east
San Juan Bautista
Parkfield-Cholame
South of Kettleman Hills
Transverse Ranges west of
Santa Barbara
West end of Garlock fault
East end of Garlock fault
Eastern San Gorgonio Pass
East and south of Salton Sea
Southern San Jacinto fault (N) Southern San Jacinto fault (S)

Mammoth Lakes-Long Valley Northern Owens Valley-White
Mountains
Off Oceanside and San Diego

Seismicity
increase

X

Adjacent
seismicity

X
X

X

X?

X
X

X

X

X
X

End of
creeping
fault

X
X

X
X

X
X
X
X
X
X

Uplift

X

Table 2

81

EARTHQUAKES 1975 -1986

WU5.7

Figure 1. Significant earthquakes (magnitude than 5.7 or greater) in California, 1975-1986. Quaternary faults from Fault Map of California compiled by Jennings (1975). Letters indicate place names used to designate earthquakes: OR, Oroville; SB, Santa Barbara; M, Mammoth Lakes-Long Valley; CL, Coyote Lake; IV, Imperial Valley; IVA, Imperial Valley aftershock; L, Livermore; V, Victoria, Mexico; E, Eureka; W, Westmorland; C, Coalinga; CA, Coalinga aftershock; MH, Morgan Hill; KH, Kettleman Hills; NFS, North Palm Springs; OC, Oceanside; CV, Chalfant Valley.

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Figure 2. Long-term creep record at Shore Road along the Calaveras fault showing the influence of the Coyote Lake and Morgan Hill earthquakes (from Schulz, 1984). Notice the absence of continued displacement (creep retardation) prior to both earthquakes.

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II6°00' Il5 e 30'

33°30' -

33°00'

32°30 -

D 1932-1940
© 1940-1950
A 1950-1960
1960-1973
* 1973-1984

1910 1920 1930 1940 I960 I960 1970 I960

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El Cenlro

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aftershocks

1930 1940 1950 I960

Time, year

1970 I960

Figure 3. (Left) Subdivisions for earthquakes in Imperial Valley; earthquake symbols indicate time interval of occurrence. (Right) Moment release with time in each subdivison. Unshaded bars represent unrelocated earthquakes originally located from intensity reports, shaded bars are relocated earthquakes, striped bars are located on the basis of immediate aftershocks, and "S" denotes swarm sequences. (From Doser and Kanamori, 1986)

33.50°

33.00°- ,

32.50°-

-116.0° -115.50° LONGITUDE

-116.00°

Figure 4. Relation of the faults and aftershock zones of the 1979 Imperial Valley and 1980 Victoria, Mexico earthquakes (solid stars) (from Silver and Masuda, 1985). Crosses and circles denote aftershocks of the 1979 event north of the border, pluses represent aftershocks south of the border. Triangles mark aftershocks of the Victoria event.

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SEISMICITY OF THE IMPERIAL VALLEY

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BRAWLEY SEISMIC ZONE

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"UNITED STATES. ^.J.- _.---- *-* ""MEXICO

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Figure 5. Seismicity in the Imperial Valley between June 1973 and November 1978 Also shown are epicenters of the 1979 Imperial Valley mainshock (TV), principal aftershock 5YiAi nSJ? ^ 1981 Westmorland earthquake (W) (large open circles) (from Johnson and Hill, 1982). Small open circle represents the epicenter of the 1940 Imperial Valley earthquake (ML=7.1); small solid circle denates epicenter of the 1942 Superstition Mountain earthquake (ML=6.5).

86

San
Jose

Location map

Halls Valley

Loma Prieta to Mount Hamilton

1973 1976 1979

Llagas to Sheep

Coyote Lake
M L = 5.9

6 August 1979

1982 1985

n

Morgan Hill
M L =6.2

24 April 1984
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Hollister

1975 1980 1985

Figure 7. (a) Relation of the 1979 Coyote Lake (dashed box) and 1984 Morgan Hill rupture zones (solid box) (from Bakun et al., 1984). Stars represent epicenters: 1, 1980 mainshock; 2, delayed aftershock; 3, 1979 mainshock. Solid triangle indicates location of the Shore Road creepmeter. (b) Length of line from Loma Prieta to Mount Hamilton with time, recent measurements are shown expanded in the inset, (c) Length of line from Llagas to Sheep. ,

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Morgan Hill
M L 6.2

24 April 1984

- *- 6 HOURS »-

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24 HOURS

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10 mm

Coyote Lak«
M -5.9
6 August 1979

«- 6 HOURS -»-

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Figure 8. Creep response at the Shore Road creepmeter from the Coyote Lake and Morgan Hill earthquakes (from Schulz, 1984). (a) Rustrak trace showing a 0.11 mm coseismic step associated with the Morgan Hill event, followed within 23 minutes by 12.9 mm of afterslip over the next 18 hours, (b) Telemetry trace showing entire sequence sampled at ten-minute intervals, (c) Rustrak trace showing a 4.2 mm coseismic step associated with the Coyote Lake earthquake, followed by 4.1 mm of afterslip in two creep events over the next 24 hours.

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DESERT HOT SPRINGS SWRRM RRER, JRN.l - MRR. 30 1985

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Figure 9. Seismicity prior to the 1986 North Palm Springs earthquake (solid circle) during the first quarter of 1985. The January 1985 swarm is represented by the dense cluster of activity just north of the original location of the 1948 Desert Hot Springs earthquake (cross). Aftershock zone of the 1986 event is oulined by the dashed lines (modified from Norrisetal., 1986)

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Figure 12. Spatial development of the Chalfant Valley sequence with time. Note especially panels A and B, indicating the earthquakes prior to the mainshock (from Cockerham and Corbett, in press). (A) 3 July to 20 July (14:29); (B) 20 July (14:29) to 21 July (14:42); (C) 21 July (14:42) to 22 July (14:42); (D) 22 July (14:42) to 31 July (07:22); (E) 31 July (07:22) to 1 August (0722); (F) 1 August (07:22) to 30 September, 1986.

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19761 1977 I 1978 I 1979 I960 1981 1982 1983 1984

Figure 13. Evidence for the strain event (shown by arrow) at Palmdale (from Savage and Gu, 1985). Plot shows gravity (top), dilatation (middle), and depth of water in a well near Palmdale as a function of time. Error bars represent one statndard deviation.

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116°40 115'

Figure 14. (a) Map of the Salton Sea area trilateration network. Stars represent epicenters of the 1981 Westmorland and 1979 Imperial Valley earthquakes (from Savage et al.,1986).

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SALTON

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Figure 15. (a) Line length L less a constant nominal length LQ with time for lines in the Salton network, southern California (from Savage et ah, 1986). Five circled measurements are considered anaomalous in that they depart by more than three standard deviations the the linear trend defined by the previous data points. Vertical lines mark time of the 1979 Imperial Valley earthquake.

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Figure 15. (b) Additional line length observations in the Salton network, arrows mark offsets associated with the 1981 Westmorland earthquake (from Savage et al.,1986).

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42

41

KNOWN OCCURRENCE OF

SURFICIAL FAULT CREEP

124 123 122 121 120 119 118 117 116 115 114

Figure 16. Known occurrence of surficial fault creep, significant earthquakes and Quaternary faults (Figure 1). Creep data for southern California from Louie et al. (1985); dashed where uncertain.

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TEN-YEAR SEISMICITY FORECAST

LIKELY LOCATIONS OF EARTHQUAKES, M * 5.7

121° 123 122

Figure 18. Ten-year forecast of seismicity. See text for discussion.

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SUMMARY OF SESSION ON CASE HISTORIES

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Summary

The session summarized here consisted of three reviews of medium-term earthquake prediction efforts in three countries: China, Japan, and the United States. The emphasis in the three reviews was different. The Chinese review was global in scope: Many dif-ferent types of precursors and different tectonic areas in China were discussed without emphasizing a particular method or region. Fewer types of precursors were mentioned in the Japanese review, and the discussion centered on the prediction attempt of the next Tokai earthquake. The U,S. paper was a study of seismicity patterns before Californian earthquakes; precursor anomalies and areas other than California were not mentioned.

If one had to pick one point that seemed most important to each of the three reviewers they might be these: In China the most important thing is to measure a large number of different parameters. In Japan the emphasis is on measuring accurately a few parameters such as crustal deformations and ground water properties, while in the U.S. paper the most important thing is new research to find out what is important.

The Chinese Review

Precursors to several large earthquakes (Mg = 7) were summarized. The reports of anomalies started typically 1000 to 2000 days before the main-shocks. The distance of the observations from the epicenter was almost never as short as 10 km, mostly larger than 100 km, sometimes as large as 1000 km. Before a main-shock typically about 50 different reports were received and approximately 10 different precursory phenomena were measured. This suggests a holistic approach in which every conceivable parameter is measured in efforts that cover large areas. The migration of seismic activity and ano-maly reports towards the main-shock epicenter received special attention.
Different models were proposed to explain early regional phenomena, and late medium-term phenomena. Through block faulting stress concentrations are created and regionally coherent seismicity patterns result. Hard and soft crustal volumes may give rise to some of the seismicity migrations, and fault creep is thought to be responsible for increases in numbers of small quakes before the main-shock.
Ma and Fu proposed that the meaning of "medium-term precursor" should be extended to include changes over periods up to 20 years. The need for this was largely based on periods of varying activity level for large earthquakes. It seems that medium-term precursors lasting 20 years make sense when the return period is several hundred to thousands of years long. In areas of 30 year return periods only the current definition of intermediate term precursor is useful. Hence it appears that there exists a need for regional definitions of what is meant by "medium-term precursor".

The Japanese Review

The Izu peninsula was instrumented to measure precursors which may allow the accurate prediction of the next Tokai earthquake. In 1978 and 1980 earthquakes of

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Hydrologic Precursors: A Critical Review

by
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U.S. Geological Survey
National Research Council Research Associate

Abstract

This review summarizes reports of anomalous flow rates or pressures of groundwater, oil, or gas that have been interpreted as earthquake precursors. Such "hydrologic" precursors have been reported, and are actively being sought, in the People's Republic of China, the USSR, Japan, and the United States. Both increases and decreases of pressure and flow rate have been observed, at distances up to several hundred kilometers from the earthquake epicenter, with precursor times ranging from less than one day to more than one year. Although information that might rule out non-tectonic causes does not appear in many published accounts of hydrologic anomalies, several recent studies have carefully evaluated the possible influences of barometric pressure, rainfall, and groundwater or oil exploitation. Several earthquakes appear to have been preceded by hydrologic anomalies for which there is no obvious explanation unrelated to the earthquake. Anomalies preceding the 1976 Tangshan, China, and the 1978 Izu-Oshima-Kinkai, Japan, earthquakes are especially well-documented and worthy of further examination.
Among hydrologic precursors, pressure head changes in confined subsurface reservoirs are those most amenable to quantitative interpretation in terms of crustal strain. The response of pressure head to earth tides determines coefficients of proportionality between pressure head and crustal strain. In particular, in a groundwater reservoir that behaves as a homogeneous isotropic porous elastic medium, pressure changes at tidal frequencies are proportional to volume strain. The same coefficients of proportionality should govern the fluid pressure response to any crustal strain field in which fluid flow in the reservoir is unimportant. Agreement between calculated and observed coseismic water level drops at Parkfield, California, is evidence that a calibration based on response to earth tides may be applied to crustal strains of tectonic origin.
Several models of earthquake generation predict precursory activity in the form of accelerating stable slip on part of the future rupture plane. If precursory slip has moment less than or equal to that of the impending earthquake, then the coseismic volume strain is an upper bound for precursory volume strain. Crustal strain can be only crudely estimated from most reported pressure head anomalies.

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magnitude 6.8 and 6.5 occurred 10 to 15 km off shore respectively. At locations 20 to 30 km distant these main-shocks were preceded by anomalies of the following parameters: Crustal uplift, earthquake swarms, volumetric strain, ground water level, ground water temperature, and Radon concentration. The durations of the anomalies ranged from a month to 1.5 years. Not all of the anomalies were equally convincing, and Shimazaki indicated that in some cases different investigators have different interpretations of the available records.

The model presented to explain the observations paid special attention to the local subduction process. Contrary to the Chinese report there was no mention of anomalies beyond 50 km. The problem of interpreting correctly the complex pattern of anomalies and come up with an accurate prediction of a large shock was presented as a very diffi-cult task.

The U.S. Review

The other two reviews asked the question: What precursory anomalies have been reported in the literature for all of China and for the Izu peninsula respectively? In con-trast the U.S. review presented an original study of seismicity patterns before all Califor-nian earthquakes with magnitudes larger than 5.6 during 1975-1986. Crustal deformation measurements in California were also considered, but no mention was made of any other precursory phenomena and of any studies outside of California. Also in contrast to the other papers no model to explain precursors was presented.

The conclusions were that future main-shocks are likely in places of increasing seismic activity, at ends of creeping fault-segments and in areas of regional crustal defor-mation.

Avoiding the discussion of precursory anomalies Wesson may reflect a sense of dis-trust of data, shared by many U.S. investigators. Omitting discussions of models for pre-cursors may reflect the wariness of interpretations without confirmed constraints.

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Nevertheless, the sizes of many anomalies within 150 kilometers of earthquake epicenters appear consistent with this upper bound. In contrast, water level anomalies at greater epicentral distances appear to be larger than this bound by several orders of magnitude.
Clearly, adequately documented water level data can yield information about the earthquake generation process. Progress hinges on better calibration of wells and on accumulation of evidence that subsurface fluids respond as expected to crustal strain.

Introduction

Unexplained changes in pressure or flow rate, color, taste, smell,
or chemical composition of surface and subsurface water, oil, and gas have come under the heading of "hydrologic" precursors. Only changes in fluid pressure or flow rate, however, are considered here. Such changes have been reported before scores of earthquakes in at least half a dozen countries. Historical reports describe "macroscopic" phenomena, such as spouting of oil from a previously dry well, that can be observed without special equipment. In the last decade, using pressure transducers and digital filtering techniques, water level anomalies only a few centimeters in amplitude have been detected. Although many scientists still doubt there is a relationship between hydrologic anomalies and subsequent earthquakes, increasing effort is being invested in water level monitoring for earthquake prediction purposes.
Among hydrologic precursors, fluid level or pressure measurements
in confined subsurface reservoirs have the best-developed theory, and the most supporting evidence, to show that they are related to crustal strain. This review will focus on water level changes, in an attempt to assess their credibility, as well as their consistency with the hypothesis that they represent response to preseismic slip on the same plane as the future earthquake. It will be apparent that many published accounts of premonitory water level anomalies are not accompanied by sufficient information to evaluate the possibility that they may have an explanation unrelated to the subsequent earthquake. Precise estimation of the crustal strain represented by a water level anomaly is also usually impossible with the available information. Nonetheless, there are some well-documented reports of credible water level precursors from which estimates of precursory strain can be inferred.

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Theoretical Effect of Crustal Strain on Water Level

Quantitative analysis of water level data requires mathematical models of two systems: first, the response of fluid pressure in the reservoir to crustal strain, and, second, the response of the observed quantity (eg., fluid level or fluid pressure in a borehole) to the changing pressure field in the reservoir.
Unless there is evidence to the contrary, it is usually assumed that the saturated rock of the fluid reservoir behaves as a porous elastic material. The general description of a porous elastic material was given by Biot (1941), and conveniently reformulated by Rice and Cleary (1976). Using the notation of Rice and Cleary (1976), the porous elastic medium can be characterized by a diffusivity, c, drained and undrained Poisson's ratios, V and V y J Skempton's coefficient, B, and a shear modulus, G. This formulation is quite general in that it allows for compressibility of the grains in the rock material, and can describe the response to arbitrary changes in the stress or strain of the reservoir.

Fluid Pressure Change Produced by Volume Strain: Confined Reservoir Earth tides or barometric disturbances of large spatial extent may
be expected to produce uniform volumetric strain in a subsurface fluid reservoir. For sources that do not induce spatial pressure gradients, it is appropriate to neglect fluid flow in the reservoir. Making this assumption in the constitutive relations given by Rice and Cleary (1976) gives

Ap = -(2GB/3)(l+v /l-2v) Ae (1)

where Ap is the change in reservoir fluid pressure and Ae is the increment of volumetric strain in the reservoir (expansion positive). For example, in a porous elastic reservoir with G = 3 GPa, B = 0.8, and V =0.3, (1) predicts 52 cm/ ye water level change. Several authors have published relationships similar to equation (1). Bredehoeft (1967) analyzed the response of a porous medium with incompressible grains. Bodvarsson (1970) also gives a relationship between reservoir pressure and volume strain, but it is not expressed in terms of the poroelastic properties of the reservoir. Van der Kamp and Gale (1983) derived an expression equivalent to (1) and showed that it reduces to the result of Bredehoeft (1967) if the grains are assumed incompressible.
The basic principle underlying the use of water wells as strainmeters is that the same ratio of fluid pressure change to reservoir volume strain should apply for earth tides, seismic events, or other tectonic processes such as earthquake precursors.

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Consequently, water wells can be calibrated as strainmeters by determining the amplitude of a single tidal constituent for which the associated volume strain amplitude is known.

Fluid Pressure Response Produced by Barometric Pressure: Confined Reservoir
The response of water level to barometric pressure is important because fluctuations of barometric pressure can produce water level changes that might be mistaken for precursors, and because the presence of diurnal and semidiurnal components in barometrically induced water level fluctuations complicate determination of the well response to earth tides. The barometric response of a confined porous elastic reservoir can be derived by assuming that the vertical stress changes by an amount equal to the change in barometric pressure, Ab , that horizontal strains vanish, and that there is no flow. An increase of barometric pressure, Ab, at the surface of the earth compresses the reservoir, an effect that by itself would cause fluid pressure to rise by an amount

Ap = (B/3)(l+v /1-v) Ab (2)

However, in an open well, an increase of barometric pressure also exerts a downward force on the fluid surface, so the net effect of an increase in barometric pressure is to cause a drop, Ah, in water level by an amount

Ah = -(l/pg)[l-(B/3)(l+V u /l-V u)]Ab (3)

where p is fluid density and g is gravitational acceleration. For the same set of material properties used in the previous section, equation (3) predicts 0.52 cm of water level drop per 1 millibar rise in barometric pressure.

Unconfined Reservoirs
Volume strain in an unconfined reservoir produces much smaller fluid level variations than in a well tapping a confined reservoir. If a volume strain, A£ , is applied to an unconfined layer of saturated thickness H, then the water table will rise by an amount

Ah =-(H/n) Ae (4)

where n is porosity (Bredehoeft, 1967). For a 100 m thick saturated interval of rock with 2% porosity, (3) predicts 0.5 cm/ye water table rise, compared with 52 cm/ye for the confined example above. Clearly, a confined reservoir should be favorable for measuring strain.

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Frequency-Dependent Response of Water Level to Reservoir Pressure Changes
Measurements of pressure in a packed-off section of well should closely track reservoir fluid pressure. However, most measurements are of water levels in open wells. For sufficiently slow changes in reservoir pressure, fluid level tracks reservoir pressure perfectly as

Ah = (l/Pg)Ap (5)

However, ability of well water level to track changes of pressure in the reservoir is dependent on frequency because any change of water level in the well requires water to flow into or out of the well. The frequency dependence is, in principle, the same for any source of reservoir pressure change. Response of water level to reservoir pressure change is influenced by the diffusivity and compressibility (or, equivalently, transmissivity and storage coefficient) of the reservoir. The radii of the well casing and of the open interval through which the reservoir is open to the well also enter the expression, since less water must flow into a narrow well to equilibrate its water level with the surrounding reservoir.
The dynamic response of a well-aquifer system was first analyzed
by Cooper et al. (1965) for the purpose of studying water level fluctuations produced by seismic waves; their analysis, which includes the effects of inertia on the motion of the water column, treats the well as a line fluid source in an infinite porous elastic reservoir. Bredehoeft (1967) illustrates the nature of the well frequency response for earth tidal periods, for which inertia is insignificant. Bredehoeft et al. (1987) have extended the work of Cooper et al. (1965) to show that the effect of a finite well radius is negligible, and to allow the well casing to have a different radius from the section of the well that is open to the reservoir. A similar analysis was carried out by Bodvarsson (1970), who derived an expression for the response of water level in a well in a confined reservoir to harmonically varying reservoir pressure by modeling the reservoir as a sphere, and the well as a spherical cavity. Bodvarsson (1970) showed that water level response falls off rapidly for fluid pressure changes at frequencies greater than a critical frequency that can range from cycles per minute to cycles per day for sandstones, down to less than a cycle per month for relatively intact crystalline rock.

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Ratio of Pressure Change to Strain as a Function of Depth

It is reasonable to assume that fluid level or pressure in a borehole responds to tectonic strain in the same way that it responds to tidal strain. However, if the fluid in the well is coming from a small number of fractures of permeability much greater than the bulk rock, it will probably be incorrect to assume that the amplitude of a tidal constituent gives the coefficient of proportionality between fluid pressure and reservoir volume strain. Such a system will be more sensitive to compression normal to the fracture plane (Hanson and Owen, 1982; Bower, 1983). In principle, it is possible to distinguish whether a single fracture is dominating the well response, or whether a uniformly porous reservoir is an adequate model by examining the amplitudes and phases of the M2 and 01 tidal constituents (Bower, 1983). However, uncertainty as to the effect of ocean loading on the tidal amplitudes and phases, and contamination of tidal frequencies by barometric or thermal disturbances of nearly equal period, contribute uncertainty to such analyses .
Whatever the mechanism of tidal response, observation suggests that deeper wells are more sensitive to tidal strain. A compilation of measured amplitudes of the M2 tidal constituent, corrected to the equator, is shown in Figure 1 as a function of well depth; the points included are listed in Table 1. Published determinations of the amplitude of the M2 tide have been excluded from Table 1 if the period of observation is less than 28 days, if the reservoir is described as unconfined, or if the amplitudes of other tidal constituents were reported but are not in approximately the correct ratios to the amplitude of the M2 constituent. Figure 1 suggests that crude bounds can be placed on the amount of crustal strain that is required to produce a given change in water pressure, provided the depth of the well is known. All points on the plot are within the region described
as

|Ah/Ael = 77(cm/)je)/km + 90 cm/ue ± 142 cm/ue (6)

Bredehoeft (1967) showed that tidal response should be approximately inversely proportional to porosity under conditions where the rock grain compressibility is negligible compared to the compressibility of the reservoir. The approximate expression is

Ah = -(K/pgn)Ae (7)

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in which K denotes the bulk modulus of water. Equation (7) can be used with porosity versus depth relations given by Magara (1980) to show that an increase of tidal sensitivity with depth is plausible; the results of this calculation are shown as the curves in Figure 1. The curves corresponding to the high-porosity estimates of the porosity-depth relation are in reasonable agreement with the estimated range of well sensitivities, while the low-porosity estimates show an increase of sensitivity with depth that is more rapid than is observed. The reason is probably that the assumption of negligible grain compressibility is poor for very low porosity rocks.

Evidence That Water Levels Measure Crustal Strain

Compared to earthquake precursors, earthquakes and creep events
are relatively well-understood tectonic sources of crustal strain. By studying the response of water levels to these strain sources, some evidence has been obtained that water wells can respond to tectonic strain in the way predicted by theory.
The North Kettleman Hills earthquake of 4 August 1985 was the first seismic event to be recorded in water wells near Parkfield, California (Roeloffs and Bredehoeft, 1985). The Kettleman Hills earthquake (ML=5.5) was located approximately 35 km from the nearest of the four water wells that recorded it. The coseismic drops in water level range from 2.1 cm to 7.8 cm in the well closest to the epicenter. The observed water level drops were corrected for the different tidal sensitivity at each well, and it was shown that these observed water level drops agreed with calculated values of elastic strain produced by the Kettleman Hills earthquake. In addition to the magnitudes of the water level drops, the time dependence of each drop was compared with the theoretical step response of a flat confined aquifer having the frequency response derived by Bredehoeft et al. (1987). For plausible values of transmissivity and storativity, the observed time dependence was consistent with the theoretical response.
Wakita (1975) studied coseismic water level changes associated with the 9 May 1974 Izu-Hanto-Oki, Japan, earthquake, and found that the pattern of rising and falling water levels approximated the quadrantal distribution of volume strain expected from the earthquake focal mechanism.
Unfortunately, these two examples in which coseismic water level changes have been successfully reconciled with estimates of coseismic strain contrast with numerous reports of coseismic water level that are much too large, or of the incorrect sense, to be explained in this way (eg., Vorhis, 1968). The occurrence of unexpectedly large coseismic water level changes suggests that foreshocks should be considered as

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possible causes of water level anomalies. Furthermore, coseismic water level changes by themselves should not be used to calibrate the strain sensitivity of water wells.
Other evidence from Parkfield that water levels respond to crustal strain has been obtained during two months when the water level record from the Gold Hill well, after removal of tidal and barometric effects, closely resembled the similarly processed record from a nearby borehole volumetric strainmeter.
There are several reports of water level fluctuations believed to
be associated with episodes of fault creep (Merifield and Lamar, 1985; Lippincott et al., 1985); all of them are from the state of California, USA. However, only at one location have propagating creep events been independently recorded by a creepmeter and a water well: several events were recorded in a well near the town of Hollister, California, located 10 m from the trace of the San Andreas fault, and 300 m from the nearest creep meter (Johnson, 1973; Johnson et al., 1973). Three subsequent studies (Wesson, 1981; Roeloffs and Rudnicki, 1985, 1986) have shown that the sizes of the water level changes associated with an event in December, 1971, can be reconciled with the response of a well in a material of plausible material properties. However, the persistent nature of the water level drop is not accounted for by the model of a steadily propagating creep event.

Effects of Barometric Pressure and Rainfall on Water Level Data

Although it is well known that water levels fluctuate in response
to changes of barometric pressure, no barometric record is provided with most reports of precursory water level anomalies. Some data, however, are presented with corrections for barometric pressure (eg., Kovach et al., 1975; Kissin et al., 1983). The simplest approach to removing the effects of barometric pressure from water level data in order to search for signals of tectonic origin is use a least-squares procedure to determine a single coefficient of proportionality between water level and barometric pressure, which is presumably an estimate of the coefficient in equation (3). This procedure is fairly, but not completely, effective. Figure 2 show water level data from the Turkey Flat at Parkfield, before and after removal of barometric pressure effects using a single coefficient of proportionality. It is clear that this procedure has not completely removed fluctuations that resemble the barometer from the data.
It is possible that the response of water level to barometric pressure is better matched by convolving the barometric pressure record with a number of coefficients. This approach has been used by Johnson et al. (1973), and by Kissin et al. (1983); however, these studies

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have not discussed the effectiveness of this procedure, nor have they reconciled the transfer function used with the theoretical frequency responses mentioned above.
Barometric efficiencies are theoretically in the range from 0 to
1, and the largest barometric variations routinely observed at Parkfield, for example, are approximately 20 millibars, corresponding to about 20 centimeters of water. Typically, barometric pressure variations have periods on the order of several days. Consequently, if water level falls several centimeters and recovers over a period of several days, barometric pressure must be evaluated as a possible cause, and a barometric record should be provided with the water level data even when an attempt has been made to filter out the barometric effects. On the other hand, a larger or more persistent water level change cannot reasonably be ascribed to barometric pressure.
The response of reservoir pressure to rainfall is often delayed by
a period depending on the thickness and permeability of the overburden, and on the distance to the reservoir outcrop. This delay can be as long as several months (eg., Lippincott et al., 1985). A threshold amount of rainfall may be required in order to initiate reservoir recharge; thus the absence of recharge in response to the first rainfall in a season is not evidence that recharge will not occur due to subsequent precipitation. Figure 3 shows data from the Flinge Flat well near Parkfield; a delay of about ten days exists between rainfall and water level rise. In addition to the delayed rise in water level due to reservoir recharge, short-period fluctuations in water level, of unknown mechanism, occur simultaneously with rainfall events. Small rises in water level at the time of precipitation have also been observed by other investigators and can sometimes be attributed to surface loading (S. Rojstaczer, personal communication).
Several years of rainfall and water level records are helpful in distinguishing real anomalies from reservoir recharge. Lack of precipitation at the time of the anomaly is not sufficient to rule out recharge as a possible cause of a water level rise. Although a long record is required, measurements once per day are adequate to study the recharge behavior of the reservoir.

The Search for Hydrologic Precursors

Water levels are being monitored for the purpose of detecting earthquake precursors in Egypt, Japan, the People's Republic of China, the Soviet Union, and the United States. Below, the observation networks are briefly described, and the role of groundwater anomalies in successful predictions is highlighted.

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United States
The most intensive water level monitoring program currently in progress in the United States is that associated with the Parkfield, California, earthquake prediction experiment being carried out under the auspices of the U.S. Geological Survey (Bakun and Lindh, 1985). It appears that nearly identical earthquakes of magnitude about 6 have occurred on the San Andreas fault near Parkfield approximately every 22 years since 1881. In anticipation that the next "characteristic" Parkfield earthquake will take place within a few years, several types of monitoring, including water levels, have been undertaken in the Parkfield area. Water levels are currently recorded at 6 wells that range in depth from 100 to 200 meters. Pressure head in each well, as well as barometric pressure and rainfall at each site, are recorded every 15 minutes and telemetered via satellite to the U.S. Geological Survey laboratory in Menlo Park, California. Other crustal deformation sensors operating at Parkfield include creepmeters, tiltmeters, borehole strainmeters, a two-color laser geodimeter network, and two leveling lines. Although the Parkfield experiment is yielding valuable information about the behavior of water wells as strainmeters, no precursory anomalies have yet been identified.
There have been several other studies of water levels in California. Lamar-Merifield Geologists, Inc., collected water level data in southern California beginning in 1976 (Lamar and Merifield, 1979; Merifield and Lamar, 1978, 1980, 1981b; Merifield, Lamar, and Bean, 1982, 1983). Over thirty wells were eventually monitored. Several were equipped with a telemetry system that transmitted data to a central observatory once per day; water levels in these wells were sampled once a minute. Other water levels were measured by hand by volunteers, at intervals ranging from once a day to once a month. Merifield and Lamar (1985) describe anomalous water level changes in several wells before two earthquakes, on 25 February 1980, and on 22 March 1982.
A 152 m deep well in the Cienega Winery near Hollister was monitored from 1971 thru 1975 (Kovach et al., 1975). Small water level minima were observed in this well before three nearby earthquakes. Three wells in the Mojave Desert, southern California, have intermittently been monitored by U.S. Geological Survey personnel since 1981 (Healy and Urban, 1985).
Since 1984, the U.S. Geological Survey has been monitoring three wells with depths from 80 to 150 m in the Long Valley Caldera, with the goal of detecting water level changes related to changes of deformation rate. Water level data are recorded at time intervals ranging from 15 to 30 minutes (S. Rojstaczer, personal communication).

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People's Republic of China
Hydrologic observations constitute an important part of the earthquake prediction data collection effort in the People's Republic of China. Observation wells in China are concentrated in the eastern half of the country, where the population density, and consequently the seismic risk, are greatest. Four hundred wells are monitored in the Beijing-Tianjin-Tangshan area, most of which are shallow, but about 30 of which tap confined groundwater, are relatively deep, and have long observation histories of at least daily water level observations (Wang et al., 1979). Many of these wells were specially drilled for the earthquake prediction program, and abandoned oil exploration wells are also utilized. Comparable numbers of wells are monitored in Szechuan and Yunnan provinces in southern China.
Among the numerous accounts of hydrologic precursors in China, three cases stand out. The 4 February 1975 Haicheng earthquake was preceded by hundreds of groundwater anomalies, which contributed to the successful short-term prediction of that event (Deng et al., 1981). Macroscopic groundwater anomalies played a role in successfully predicting the 1976 Sungpan-Pingwu earthquake, in spite of the non-occurrence of foreshocks (Wallace and Teng, 1981). The best documented hydrologic precursors in China were those that preceded the disastrous 28 July 1976 Tangshan earthquake, for which no short-term prediction was issued. These anomalies have been carefully examined by Wang et al. (1979) and by Wu et al. (1984), and while some of them appear to have been influenced by rainfall or by groundwater exploitation, others do not have simple explanations unrelated to the earthquake. Moreover, other types of information are available: coseismic water level changes, credible precursory resistivity changes (Madden, 1987), and pre- and post-seismic leveling data (Zhang, 1982). This case clearly merits thorough study.

USSR

In the Soviet Union, earthquake prediction observations are grouped into "polygons", in which a variety of different measurements are made. There are at least eight of these polygons, located in the seismically active regions of the Caucasus, Central Asia, Kamchatka, and the Kuril Islands. Water level or flow measurements are made in all these regions -(Sidorenko et al., 1979).
The time and magnitude of an earthquake in the Pamir Mountains on
1 November 1978 were successfully predicted on the basis of many kinds of anomalies, including groundwater anomalies in the Fergana Valley polygon, Uzbek SSR (Simpson, 1979; Asimov et al., 1979).

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Japan
Although water level monitoring was not initially believed to be well-enough understood to justify its inclusion in the Japanese earthquake prediction program, that point of view has changed. Water levels are now monitored in 11 deep wells as part of the effort to make a short-term prediction of the expected Tokai earthquake (Earthquake Engineering Research Institute, 1984).
The best documented occurrences of precursory water level changes
in Japan were those that preceded the 14 January 1978 Izu-Oshima-Kinkai earthquake (Ms = 6.8) (Wakita, 1978; Oki and Hiraga, 1978, 1987), although no short-term prediction was issued for this event. Water level anomalies were recorded in three deep observation wells, and at over 30 shallower wells by a primarily volunteer organization known as the Catfish Club. In addition to the water level anomalies, there appear to have been precursory changes in water temperature and radon concentration. Anomalous deformation was recorded by borehole volume strainmeters and surface geodetic measurements. Re-examination of the hydrologic anomalies preceding this earthquake is underway, including study of the effects of groundwater pumping on the anomalies, and of the uniqueness of the anomalies in long observation records (Shimazaki, 1987).

Several wells have been monitored since 1985 in a search for precursors to .earthquakes near Aswan Dam in Egypt (D. W. Simpson, personal communication).

Characteristic Features of Hydrologic Precursors:
Some Hypotheses

Observers of hydrologic anomalies have suggested a number of features they believe to be common characteristics of these precursors. These hypotheses are summarized below.

Distance from epicenter
A remarkable feature of the case histories listed in the Appendix
is that precursory groundwater anomalies are believed to have been observed at distances of several hundred kilometers from the earthquake epicenter. Monakhov et al. (1983) and Sadovsky et al. (1979) state that small "rebound-type" anomalies can be observed up to 100 km away for events of magnitude less than 4, and up to 900 to 1000 km away for events of magnitude 7 or 8. Before the 1976 Tangshan earthquake, potentially premonitory groundwater anomalies were observed as much as 200 km from the epicenter (Wang et al., 1979). Anomalies at distances

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between 250 and 360 km from the epicenter were reported for the 1983 Heze and 1982 Lulong earthquakes in China (Wang and Wu, 1986). Groundwater anomalies 150 km from the epicenter have been reported for three successfully predicted earthquakes: 1975 Haicheng, China (Deng et al., 1981), 1976 Sungpan-Pingwu, China (Wallace and Teng, 1980), and 1978 Pamir Mountains, USSR (Simpson, 1979).

Sensitive Locations
Particularly in the People's Republic of China, it is believed that wells drilled into fault zones, and ideally, at intersections of two or more fault zones, are especially sensitive to premonitory groundwater variations (eg., Zhu and Zhong, 1979; Wu and Wang, 1986). For example, before the 1976 Sungpan-Pingwu earthquakes, groundwater and other "macroscopic" anomalies appeared to approach the epicentral area along the Lungmenshan fracture zone (Wallace and Teng, 1980). Some concentration of groundwater anomalies along active faults was reported prior to the 1975 Haicheng earthquake (Deng et al., 1981). The more distant groundwater anomalies preceding the 1976 Tangshan earthquake were observed along faults (Wang et al., 1979). In the United States, the precursory water level anomalies reported by Kovach et al. (1975) were observed in a well drilled into the San Andreas fault zone.
Wells drilled into fault zones may be more likely sites of precursory anomalies because the fault is directly involved in the generation of the impending earthquake, or because local faults that slip in the intensifying stress field around the future epicenter concentrate strain in their vicinities. On the other hand, it may be that fault zone materials respond with greater water pressure changes to relatively homogeneous preseismic strain. The first two possibilities could be distinguished from the third by studying the response to earth tides of "sensitive" wells in fault zones.

Sense of Precursory Water Level Changes
Although the case histories listed in the Appendix make it clear that both rising and falling water levels have been interpreted as precursors, many authors have expressed the view that falling water levels predominate. According to Monakhov et al. (1983) and Sadovsky et al. (1979), the most common form of precursor is a several centimeter drop several days before the earthquake, which is beginning to recover as the earthquake occurs; this type of anomaly is often referred to as a "rebound" anomaly. Anomalies having the same form have also been reported to occur over longer time scales (eg., Kovach, 1975). Wang (1985) also concludes that water levels commonly drop before earthquakes. These water level drops are believed to be related to increase of porosity and permeability due to fracturing, with the

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subsequent recovery attributable either to influx of fluid or to compression.
Nonetheless, there are many reports of rising water levels that have been interpreted as precursors. Deng et al. (1981) state that many more rising than falling water levels were observed before the 1975 Haicheng, China earthquake. Both rising and falling water levels would be expected if the anomalies represent the response to volumetric strain produced by fault slip. The possibility that aseismic precursory slip is the source of water level anomalies is discussed more fully below.

Time Migration
Researchers differ on where groundwater anomalies will first be observed. Wang and Li (1979) reviewed water level anomalies before the Haicheng, Tangshan, and Sungpan-Pingwu earthquakes, and concluded that anomalies first appeared at large distances from the epicenter, and then migrated toward the epicenter, particularly along the direction of future fault rupture, or along fault lines parallel to that direction. This pattern is discussed in detail for the 1975 Haicheng, China earthquake by Zhu and Zhong (1979), although Deng et al. (1981) describe a more complex time-space pattern for the same set of anomalies. For both the Haicheng and Sungpan-Pingwu earthquakes, the migration was of "macroscopic" anomalies and occurred over a distance of 150 to 200 km and a time period of about 3 months. The explanation for the converging of anomalies toward the epicenter is the gradual concentration of stress in the epicentral area, which is believed to be the last to fail because it is a region of relatively high strength (Wang and Li, 1979).
On the other hand, Sadovsky et al. (1979) believe that water level anomalies begin sooner, and are larger, close to the epicentral area than further away. Because the anomalies discussed in Sadovsky et al. (1979) occur several days before earthquakes 500 or more km away, the time and distance scales of this migration are not comparable to the migration of anomalies before the Haicheng and Sungpan-Pingwu earthquakes. A case that can be directly contrasted with the Chinese examples is the 1978 Izu-Oshima-Kinkai earthquake in Japan. Anomalies appeared 3 to 5 weeks before this earthquake within 50 km of the epicenter, but only several days before the earthquake at distances between 50 and 150 km (Oki and Hiraga, 1979).

Water Level Anomalies and Preseismic Slip

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The new information we hope to measure with water wells is aseismic precursory strain. Very little is known about how aseismic strain occurring in a future hypocentral volume is distributed in space and time, but one class of theories suggests that the precursor will take the form of slip over a portion of the fault that will rupture in the future earthquake. If the material around the fault is assumed to remain elastic, then the volume strain produced by such slip can easily be calculated. The moment of precursory slip is unknown, but the moment of the future earthquake may plausibly be taken to be an upper bound. A crude method of estimating volume strain from water level change is to use the relationship suggested by the plot of tidal sensitivity versus depth (Figure 1 and equation (6)). If the anomaly is attributable to precursory slip of moment less than that of the future earthquake, then the coseismic volume strain should exceed the strain calculated from the water level change. In particular, the strain obtained from the water level change should lie below the curve of coseismic volume strain versus distance along the azimuth of largest coseismic volume strain for a particular fault orientation. This hypothesis can be tested for earthquakes of different moments by plotting strain per unit moment, as is done in Figure 4 for the data set listed in Table 2. The data set consists of water level anomalies for which well depth, epicentral distance, earthquake magnitude, and size of anomaly were reported. The vertical error bars include the uncertainty in the tidal calibration from equation (6), as well as a factor of 10 uncertainty in the seismic moment. Although no horizontal error bars are plotted in Figure 4, an estimate of rupture length would be an appropriate measure of the uncertainty in distance. Empirical relationships between magnitude and rupture length (eg., Schwartz et al., 1984; Wyss, 1979) suggest that for all but two of the earthquakes listed in Table 2, the rupture length is less than 10 km. The two exceptions are the 1976 Tangshan earthquake, for which a rupture length of 140 km is reported by Butler (1979), and the 1978 Izu-Oshima-Kinkai earthquake, for which a rupture length of 17 km is reported by Shimazaki and Somerville (1978).
The comparison made in Figure 4 must be viewed with caution, because the conversion of water level to strain is crude, because a more extensive data set should be used, and because the data have not been corrected for azimuth relative to the strike of the earthquake fault. Bearing these inadequacies in mind, Figure 4 suggests that, out to 150 km from the epicenter, many reported anomalies can be accounted for by slip of moment less than or equal to the future earthquake. In particular, the sizes of the water level anomalies preceding the 1976 Tangshan and 1978 Izu-Oshima-Kinkai earthquakes can be accounted for in this way. Further than 150 km from the epicenter, most reported anomalies appear to correspond to an amount of strain greater than the

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coseismic strain, even allowing for the considerable uncertainty in seismic moment, well sensitivity, and distance from the well to the initiation point of the rupture.
Although no firm theoretical basis exists for requiring preseismic slip to be less than coseismic slip, many workers believe that preseismic slip is probably only a small fraction of coseismic slip. This belief is supported by data from borehole strainmeters during several moderate earthquakes, including the 1978 Izu-Oshima-Kinkai earthquake, that show no evidence of preseismic slip, and consequently suggest that the moment of preseismic slip must be less than a few percent of the seismic moment of the earthquake (Johnston et al., 1986) Dieterich (1986a,b) has studied behavior of spring-slider models embodying a slip rate and rate-history dependent frictional constitutive law. In such a system, slip instability is always preceded by accelerating stable slip over a patch of the fault, which might give rise to premonitory signals in the interval from 10 days to 10 minutes before an earthquake. The moment of premonitory slip is governed by a characteristic slip distance and is not necessarily related to the moment of the impending earthquake, which may rupture an area larger than the nucleation patch.
The characteristic slip distance is believed to be related to the scale of roughness of the fault surface (eg., Tullis, 1987). Estimates of this distance from laboratory measurements are less than 50 microns, but larger values may be appropriate for earthquake faults. If laboratory values of characteristic distance are appropriate, then only a fraction of a millimeter of precursory slip would be expected over a radius of several meters. However, if the characteristic distance is as large as 50 mm, then precursory accelerating slip of 250 mm could occur over a patch of radius several km, yielding precursory slip with a moment of 5.5 x 10 dyne-cm (Dieterich, 1986a). This moment is not large enough to account for very many of the anomalies listed in Table 2, even when only anomalies with precursor times less than 10 days are considered.
Models based on slip rate and rate-history dependent friction laws predict that once slip has begun to accelerate over the nucleation patch, displacement on that patch increases very rapidly. To a network of water level sensors distributed around the fault, each with a non-zero strain detection threshold, this increasing displacement will be detected at increasingly greater distance from the nucleation patch. Consequently, the anomaly would appear to emanate from the future epicenter. Anomalies fitting this theory are those that increase monotonically in a sense consistent with the future coseismic volume strain, beginning within 10 days before the earthquake. A possible example is the accelerating water level drop at Funabara preceding the 14 January 1978 Izu-Oshima-Kinkai earthquake in Japan (Wakita, 1981),

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although this anomaly began about one month before the earthquake. Neverthless, it is clear that many reported anomalies do not have this time dependence.
Stuart et al. (1985) propose a model of the earthquake generation process in which the fault zone, taken to be a vertical plane in an elastic half-space, has spatially varying resistance to frictional slip. In particular, a patch of the future earthquake rupture area is assumed to obey a strain softening relation between slip and frictional resistance once a peak stress has been exceeded. Rate and rate-history dependent friction laws reduce to this relationship for large values of the characteristic slip and a constant slip velocity. Outside this patch, the fault slips freely at a constant stress, or is completely locked. When such a system is subjected to steadily increasing tectonic shear stress, four distinct stages of behavior result. During the first stage, the patch behaves similarly to other areas of the fault. Its increasing resistance to frictional sliding begins to become significant at the beginning of the second, "slow load" stage, but during this stage, all points on the patch are still in a state of strain hardening. Increasing tectonic load and concentration of stress on the patch gradually bring its outer edges past their peak stress, at which time the third, "precursory" stage of accelerating slip begins. Earthquake instability results when tectonic loading and stress concentration combine to load the central, unfailed, portion of the patch more quickly stress can be relieved with any finite rate of slip. The stage at which accelerating, aseismic fault slip might be observed is the third, "precursory", stage. For the model parameters obtained by calibrating the model to agree with observed creep and line-length changes at Parkfield, CA, Stuart et al. (1985) suggest that these changes might become observable about one year before the predicted moderate earthquake. The length of the precursor time in other locations would be expected to depend on the tectonic loading rate, as well as on the slope of the unloading portion of the slip versus frictional resistance relation; clearly, there are few other situations where these quantities can be constrained even as well as they are at Parkfield.
Accelerating slip occurs first at the edges of the patch and spreads inward toward the future hypocenter. The creep and line-length changes at Parkfield are most consistent with a patch radius of 3 km, which is half to a third the size of the rupture area of the 1966 Parkfield earthquake. In this situation, although the locus of accelerating slip converges toward the hypocenter, it does so over a small enough distance scale that the inward migration might not be apparent in water level records. Unless the patch dimension is very large for large earthquakes, this mechanism cannot explain the movement of hydrologic anomalies toward the epicenter over distances on the

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order of 100 kilometers, such as was observed before the 1976 Sungpan-Pingwu earthquakes.

Conclusions

There is clearly a solid theoretical basis for expecting water level anomalies if there is precursory aseismic deformation producing volume strain in the crust. Increasing evidence shows that water wells can behave as calibrated strainmeters.
We11-documented premonitory groundwater anomalies have been reported that are long-term (eg., Tangshan, 1976), intermediate-term (eg., Haicheng, 1975; Izu-Oshima, 1978), and short-term (eg., Pamirs,1978) precursors. Although many investigators state that falling water levels are more common, both rising and falling water levels have been reported, and are consistent with regions of contraction and extension, respectively, in the crust. Anomalies occur as far as several hundred kilometers from earthquake epicenters, and are often reported to be concentrated in fault zones. Migration both toward and away from the epicenter has been observed.
Real progress in the use of water levels to measure crustal strain requires better documentation of the data. Published accounts of hydrologic precursors should include information on well depth, length of observation, measurement technique, and rainfall response of each well. The possible role of barometric pressure must be evaluated for water-level anomalies that are 20 cm or less in amplitude and of duration of the order of several days; simple filtering techniques for removing barometric pressure are not always sufficiently effective. Attempts should be made to express water level changes in terms of crustal strain. Although all reports of water level anomalies to date are deficient in some aspect, the cases of the 1976 Tangshan, China, and 1978 Izu-Oshima-Kinkai, Japan, earthquakes are the best documented and should be thoroughly studied.
Frequent sampling was not needed to record several credible hydrologic precursors. Measurements of water levels on a daily basis with a tape are clearly adequate for discerning long-term changes and can provide baseline data for more detailed future monitoring. On the other hand, attempts should be made to evaluate wells as calibrated strainmeters by studying their response to earth tides. The required data could be collected using a portable recorder rotated among the wells in an observation network, remaining at each well for about two months, in situations where the cost of continuously recording such information at every well is prohibitive.
Figure 4 suggests that many water level anomalies within 150 km of earthquake epicenters are not too large to be explained by preseismic

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slip of moment comparable to the earthquake. Although this conclusion must be regarded as tentative, it appears quite likely that water level anomalies can help answer the question of the moment and timing of accelerated preseismic slip, if it occurs.

Appendix
Reported premonitory variations in level or flow of water, oil, or gas

Following is a brief summary of the available information on reported premonitory variations in the level or flow of water, oil, or gas, obtained primarily from English language literature. Supporting information on well depth, barometric pressure, rainfall, and length of the period of observation is critical in evaluating the possible relation between any reported anomaly and the earthquake preparation process. Thus, such availability is noted in the descriptions below.

The list of reports is in chronological order. The word "magnitude" is used to denote value on an unspecified magnitude scale. An attempt has been made to refer to Chinese place names using the current official Mandarin transliteration; other transliterations under which information appears are included in parentheses.

18 January 1964, magnitude 6.7, Chia-yi, Taiwan: Tubing pressure in oil-field wells increased between 15% and 60%, beginning on 9 January (Wang and Li, 1975).

8 March 1966, magnitude 6.8, and 22 March 1966, magnitude not stated, Xingtai (Hsingtai), Hopei, China: groundwater levels rose or fell within a 230 km by 190 km area before the first event, and over a 300 km by 220 km area before the second event. (Wang and Li, 1979) Scientists measuring well water levels with tapes in the epicentral region after the first event observed drops in water level before aftershocks (Zhang Y.-Z., personal communication).

3 December 1970, magnitude 5.5, Hai-yuan, Ningsia, China: Water level rises of as much as 0.8 m were observed (Wang and Li, 1979).

24 February 1972, ML = 5.0, Bear Valley, California, USA: In a 152 m deep well, 25 km from the epicenter, a water level drop of approximately 5 cm occurred about 60 days before this earthquake and persisted for about 25 days. The water level drop does not appear to

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be due to the seasonal effect of rainfall. A correction has been applied for barometric pressure variations. This well was monitored beginning in May, 1971 (Kovach et al., 1975).

4 September 1972, ML = 4.7, Stone Canyon, California, USA: In the same well for which the precursor of the 24 February Bear Valley event was reported, a water level drop of about 8 cm was observed about 50 days before the earthquake, with a duration of about 40 days. This earthquake occurred 17 km from the well (Kovach et al., 1975).

15 January 1973, ML =4.0, Lewis Ranch, California, USA: In the same well for which the precursor of the 24 February Bear Valley event was reported, there was a 5 cm water level drop 40 days before the Lewis Ranch earthquake, which persisted until the earthquake occurred. The epicenter of the Lewis Ranch earthquake was 10 km from the well (Kovach et al., 1975).

23 April 1974, magnitude 5.5, Li-yang, Kiangsu, China: rapid water level rises occurred in two wells (one rise was about 1 meter); water overflowed from a third well (Wang and Li, 1979)

4 February 1975, magnitude 7.3, Haicheng, China: About 240 water level anomalies were reported between 16 November, 1974, and the time of this earthquake. These anomalies are described in detail by Zhu and Zhong (1979) and Deng et al. (1981).

8 April 1976, Ms = 7.1, and 17 May 1976, Ms = 7.1, Gazli, USSR: There was a sharp drop in water level (Sadovsky et al., 1979).

28 July 1976, magnitude 7.8, Tangshan, China: Numerous water level anomalies were observed, perhaps as much as 6 years before this earthquake. They are described in detail in Wang et al. (1979), Wang and Li (1979), Zhang and Qiuo (1979), and Wu et al. (1984). Wang et al. (1979) and Wu et al. (1984) discuss the effects of rainfall and groundwater exploitation on the data.

16 August 1976, magnitude 7.2, Sungpan-Pingwu, Szechuan province, China: 395 groundwater changes were observed beginning in March, 1976, becoming larger in June 1976, and clustering along the fault in the days before the earthquake. (Wang and Li, 1979; Wallace and Teng, 1980; Rikitake, 1982; Wang, 1985)

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24 November 1976, Ms - 7.3, Caldiran, Van province, Turkey: About 1 day before the earthquake, increased discharge from a spring, accompanied by oil seepage, occurred at the eastern end of the 55 km long section of fault along which surface displacement was mapped (Toksoz, 1979).

12 May 1977, magnitude 6.5, Ningho, Hopei, China: a usually dry well in Chinghsien county, 110 km from the epicenter, discharged oil and gas between 27 March and 12 April, 1977; oil production in this well prior to the 28 July 1976 Tangshan earthquake was also believed to have been a precursor. Similarly, a well in Jenchiu county, 160 km from the epicenter, produced water from 10 March to 12 March, and gushed oil from 22 March to 24 March; it had also gushed oil before the 1976 Tangshan earthquake (Wang and Li, 1979).

23 July 1977, magnitude 5.5, Xingjiang ("Sinkiang"), China: A 380 m deep well, 50 km from the epicenter, began to produce four times its normal amount of fluid , with double the normal content of oil, beginning on 10 July, 1977. The same well began to produce only water and mud on 21 July, and stopped flowing after the earthquake. Flow rates increased in two other wells on 19 July, and another previously dry well yielded gas on 23 July (Wang and Li, 1979).

14 January 1978, Ms = 6.8, Izu-Oshima-Kinkai, Japan ("West Off Izu-Oshima earthquake"): Three water level anomalies were observed in deep monitoring wells. Water level in a 500 m deep well at Nakaizu, 30 km from the epicenter, fell approximately 30 cm at the beginning of December, 1977, and had risen again by about 20 cm by the time the earthquake occurred. In a 600 m deep well at Funabara, a rising trend of water level reversed itself in the middle of December, 1977; water level fell about 1 m before the earthquake. Funabara is 35 km from the epicenter. 90 km from the epicenter at Omaezaki, water level in a 500 m deep well dropped by 25 cm, beginning on 28 December, 1977, and then began to rise on 5 January, 1978. These three water level anomalies are described by Wakita (1978), who refers to more detailed reports in the Japanese literature. In addition to the deep-well anomalies, water level anomalies were observed at over 30 locations, mostly shallow wells, at which groundwater levels were being monitored by the Catfish Club. These anomalies are described in Oki and Hiraga (1978), in a very complete report that includes information on the response of the monitoring wells to barometric pressure, rainfall, and groundwater pump ing.

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27 February 1978, magnitude 4.6, and 7 September 1978, magnitude 4.5, Ashkhabad, USSR: About two months prior to the first of these events, water level in a well 120 km from the epicenter dropped about 1 meter, and had recovered by about half that amount when the earthquake occurred. In the same well, a drop of about 1.5 m began 5 months before the second event, and was beginning to recover when the earthquake occurred, at a distance of 20 km from the well. Water level observations had been made in this well since 1970 (Asimov et al., 1979).

22-23 March 1978, sequence with magnitudes up to 8, Kuril'sk, USSR: In a well at Goryachiye Kluchi, 170 km from the epicentral region, flow of water increased from about 1100 to over 1200 I/hour 4 days before the first magnitude 7 event; the flow dropped to its previous rate 1 day before the earthquake. A day or two after this event, the flow again decreased, and a magnitude 8 event occurred several days later. It is stated that atmospheric pressure and precipitation do not affect the flow rate in this well (Monakhov et al., 1983).

1 November 1978, magnitude 7, Pamirs, USSR: Six hours before this event occurred, its time was successfully predicted to within a few hours, and its size was predicted to within 1/2 of a magnitude unit. The location, however, was predicted only to within a few hundred kilometers. A long-term decrease in water levels near Khodzhaabad, in the Fergana Valley, beginning the previous July, was a possible intermediate-term hydrologic precursor. In the same area, cessation of flow in the deepest of 11 wells being monitored, and decrease of flow in the remaining wells, at the end of October, was interpreted as a short-term precursor. These hydrologic precursors occurred at a distance of 150 to 200 km .from the epicenter (Asimov et al., 1979; D. W. Simpson, 1979).

7 July 1979, Ms = 6.0, Liyang, China: Of six wells being monitored within 200 km of the epicenter, 2 wells about 370 m deep showed several centimeter rises in water level in the 2 weeks before the event. The water level rise was observed in a well 45 km from the epicenter 13 days before the earthquake; it was observed in a well 130 km from the epicenter 5 days before the earthquake. Another well, 600 m deep, showed a drop in flow rate 5 months before the even; this well was 60 km from the epicenter (Wang and Wu, 1986).

25 February 1980, ML 5.5, southern California, USA: Two wells located 31-32 km from the epicenter showed anomalous rises of water level lasting about 4 hours, 4 days before the earthquake. The rise totaled 45 cm in a 32 m deep well, and 2.3 cm in a 130 m deep well (Merifield

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and Lamar, 1985).

11 December 1980, magnitude 5.1, Nazarbek, USSR: Between 30 and 90 days before this event, a 5 cm drop in water level was observed in a 235 m deep borehole at Kirn, 170 km from the epicenter; this drop persisted until about 2 months after the earthquake. In the same borehole an additional 0.5 cm drop occurred 5 days before the earthquake, and recovered after 3 days. Two days before the earthquake, there was a 3.3 cm drop in a 305 m deep well at Asht, which is located 150 km from the epicenter. Water level observation in these wells began in early 1980; their barometric response and their response to rainfall have been studied. Tidal signals have been observed at Kirn, but do not appear to occur at Asht (Kissin et al., 1983).

13 August 1981, Ms = 5.8, Fengzhen, China: a water level drop of one centimeter was observed in a 217 m deep well situated 113 km from the epicenter 18 days before the event; water level rose in the two days before the earthquake (Wang and Wu,1986).

9 November 1981, Ms = 5.8, Lunyao, China: of 9 wells being observed within 200 km of the epicenter, 2 artesian wells showed changes of water pressure on the order of 4 to 20 cm of water in amplitude. The first anomaly occurred 77 days before the earthquake in a 2694 m deep well 150 km from the epicenter, and the second occurred 11 days before the earthquake in a 3034 m deep well 67 km from the epicenter (Wang and Wu, 1986).

22 March 1982, ML = 4.5, southern California, USA: Rapid fluctuations of water level occurred in a 133 m deep well 13 km from the epicenter. These fluctuations occurred over a 4 day period and are tentatively attributed to gas bubbling (Merifield and Lamar, 1985).

3 July 1982, Ms = 5.4, Jianchuan, China: In a well of unspecified depth 10 km from the epicenter, water level fell 13 cm 13 days before the earthquake (Wang and Wu, 1986).

19 October 1982, Ms = 5.3, Lulong, China: Three of the five wells within 300 km of the epicenter at which water levels were being observed showed water level rises of several centimeters. The change was observed 7 days before the earthquake in a well 300 km away, and 2 days before the earthquake in wells 250 and 34 km away. Well depths range from 300 to 2700 m. In the well 300 km from the epicenter, water level had been monitored for at least the previous year (Wang and Wu, 1986).

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3 March 1983, Ms = 5.4, Xingjiang, China: Two wells 11 and 12 km from the epicenter showed water level changes of 13 and 20 cm, respectively, two days before the earthquake (Wang and Wu, 1986).

7 November 1983, magnitude 5.9, Heze, China: Of 58 wells within 400 km, 8 wells, with depths between 1000 and 2700 m, showed anomalies of water level, oil production, and flow. In two wells showing water level changes, the change occurred 27 days before the earthquake in a well 80 km from the epicenter, and occurred 13 days before the earthquake in a well 360 km from the epicenter (Wang and Wu, 1986).

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Table 1. Locations of wells for which M2 amplitude is plotted in Figure 1. All amplitudes listed in the table have been reduced to the Equator.

Location Latitude, Depth, M2, Reference degrees meters cm

Kiabukwa , Zaire
Carlsbad, New Mexico, USA
Iowa City, Iowa, USA
Duchov, Czechoslovakia
Turnhout, Belgium
Heibaart, Belgium

- 7.78
32.30
41.65
50.62
51.32
51.38

2400
86
252
600
2175
1660

7.67 Melchior(1983)
0.63
2.06
3.45
3.79
2.93

Gaocun, Tianjin, China
Fanxian, Fanxian, China
Wall, Beijing, China

Gold Hill, California, USA Flinge Flat, California, USA Turkey Flat, California, USA

Crystallaire, California, USA

Chalk River 3, Ontario, Canada Chalk River 8, Ontario, Canada

Marysville, Montana, USA

CB136, Virginia, USA

3402 8.13 Liu & Zheng
2267 2.73 (1985)
657 4.28 (see note 1)

35.83 88 0.78 Roeloffs
35.93 122 0.37 & Bredehoeft,
35.89 177 0.93 unpublished

34.47 853 6.87 Bredehoeft
et al.(1986)

46.01 161 4.29 Bower(1983)
46.01 300 3.52

46.50 2000 4.85 Narasimhan
et al. (1984)

37.13 136 2.14 Rhoads &
Robinson (1979)

Note 1. Tidal amplification is stated in this reference and was used to calculate the amplitude of the M2 constituent.

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Table 2, continued.

Earthquake
Location

Date

(YYMMDD)

Magni-
tude
(note 1)

Seismic
moment

(note 2)

delta
(km)

well
depth
(m)

dh
(cm)

Reference

Tangshan, China 760728 7.7 Ms 1.8E27 77 (note 4) 63
104
110
162

60
61
100
63
283

60.0 Wu et al.
- 120. (1984)
260.
- 73.
- 70.

California, USA 800225 5.5 1.0E24 31 32 45.0 Merifield
and Lamar
(1985)

Notes. (1) magnitude scale omitted where unknown.
(2) unless otherwise indicated, seismic moment was estimated
from the moment-magnitude relation given by Hanks and Boore (1984), Figure 2, for California earthquakes, or from Nuttli (1985), equations 6a,b,c and 7a,b,c, for earthquakes not in California. Magnitudes of unspecified scale were assumed to be mb.
(3) Seismic moment from Shimazaki and Somerville (1978).
(4) Seismic moment from Butler et al. (1979).

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M2 RMPLITUDE REDUCED TO EQUflTOR

o 2 4 6 8 10 12

flmplltude, cm of water

Figure 1. Amplitude of M2 tidal constituent versus well depth for the wells listed in Table 1 (symbols). The shaded region rep-resents the relationship given in equation (6). The curves are estimates of M2 amplitude as a function of depth obtained from equation (7) and porosity-depth relations given by Magara (1980).

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E
O

PRECURSORY WRTER LEVEL PLUCTURTIONS

RISING
PRLLIN6

CD
o

300 350 400

Well - Epicenter distance, km

Figure 4. Comparison of strain represented by water level anomalies with coseismic volume strain. Symbols: strain represented by the water level anomalies listed in Table 2, converted to strain using equation (6), and divided by the seismic moment of the subsequent earthquake. Curves: Absolute value of volume strain as a function of distance along the azimuth of maximum volume strain, for several source mechanisms: vertical strike slip, at surface and 9 km deep, 45 degree dip slip, at surface and 9 km deep, and 20 degree dip slip, 20 km deep. Curves represent strain per dyne-cm of moment, assuming a shear modulus of 3 x 10 11 dynes/cm 2.

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INTERMEDIATE PERIOD RESPONSE OF WATER WELLS TO CRUSTAL STRAIN: SENSITIVITY AND NOISE LEVEL

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ABSTRACT

The response of water wells to earth tides indicates that they can be sensitive to small crustal strains. Groundwater flow induced by the presence of the water table can significantly attenuate this sensitivity. The attenuation of strain sensitivity as a function of frequency can be inferred from the response of water wells to atmospheric loading. For the wells examined in this study, significant attenuation due to water table effects can occur when strains accumulate gradually over periods of days to weeks. Even if this attenuation is present, however, wells can still be used as accurate strain meters over this range in period. In the frequency band of 0.025 to 2.5 cycles/day, the noise level of the raw water level records examined increases 25 db per decade in frequency. Much of this noise is due to the influence of atmospheric pressure. When the effects of atmospheric pressure are removed from the record, the noise level is reduced to 20 db per decade for frequencies above 0.08 cycles/day, a rate typical of high quality strain meters. Below a frequency of 0.08 cycles/day, the water level records show a noise increase of roughly 35 db per decade, a change which may reflect the influence of precipitation. For periods slightly less than a month, the two wells examined possess a higher accuracy than can be obtained from the best geodetic distance measurements.

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INTRODUCTION

Although it is well known that water wells respond systematically to small strains induced by earth tides and atmospheric pressure fluctuations (Jacob, 1940; Bredehoeft, 1967; Van der Kamp and Gale, 1983), the use of water wells for the purpose of monitoring tectonically induced strain has been largely qualitative. There have been very few successful attempts to calibrate water level fluctuations in response to known tectonic events in terms of strain and/or compare a water well's response to other strain measurements. Sterling and Smets (1971) include a power spectrum from a water well's response to a seismic event which indicates that in the frequency band of 10 4 to 10 2 Hz a well can be highly sensitive to dilatation. Johnson et al. (1973, 1974) monitored water levels from 1971 to 1973 in a well located near the Almaden Winery within 10 m of the surface trace of the San Andreas fault; they found that water level fluctuations of several centimeters which lasted from hours to several days were associated with creep events. Based upon the response of their well to atmospheric loading, Johnson et al. were also able to infer the mean stress at the well induced by the associated creep events. Wesson (1981) and Roeloffs and Rudnicki (1984) have analyzed one of the water level - creep events observed by Johnson et al. and have found that the water level fluctuation is qualitatively consistent with the magnitude of the creep event. The same well observed by Johnson et al. showed correlations with two creep events during 1975-1976 (Mortensen et al.. 1977); water level fluctuations of 3 centimeters were associated with creep induced strains of approximately 10" 7.

The response of the water well near the Almaden Winery suggests that water wells can measure tectonic strain. There are, however, some inherent problems with using water wells as strain meters. These problems can be separated into two broad classes: those that are strictly related to the bulk material properties of the rock or sediment in direct communication with the well; those that are due primarily to groundwater flow.
The principal problem related to bulk material properties is that the sensitivity of a well to strain is highly dependent on the porosity and

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elastic properties of the rock or sediment with which it is in hydraulic communication. Wells in communication with rock or sediment which is highly compressible and porous cannot be expected to be highly sensitive to tectonic strain. These points are discussed in a related paper (Rojstaczer, 1987).
Many other problems with using water wells as strain meters are due primarily to the influence of groundwater flow. The principal problem related to groundwater flow is that the water level in a well responds to hydrology as well as strain. A year of above average rainfall will likely cause a low frequency rise in water level; conversely, a year of below average rainfall will likely cause a low frequency drop in water level. The hydrologic influences of rainfall can be viewed as noise placed upon any strain signal which might be present in the water well record and interpretation of changes in water level as being solely the result of strain will have some inherent error.
Groundwater flow can cause another significant problem inherent in the
use of water wells as strain meters; it may reduce strain sensitivity, and restrict the usefulness of water wells to a narrow frequency band. Figure 1 shows an idealized schematic of a well in communication with saturated rock undergoing strain and indicates three potential ways for groundwater flow to reduce strain sensitivity. The first source is the limited hydraulic communication between the well and the saturated rock; if the well is to be a gage of pore pressure, changes in pore pressure must be accompanied by flow into or out of the borehole. For strains above some limiting frequency (which depends on the borehole geometry and the material properties of the saturated rock), groundwater flow will be too slow to allow the full pore pressure signal to be seen in the well. Fortunately, this can usually be mitigated by placing a packer beneath the water surface so that only small volumes of fluid must move in and out of the borehole to change fluid pressure within the well.
The second source of sensitivity reduction due to groundwater flow is flow from (in the case of strain being compressional) or to (in the case of strain being extensional) the strain induced pore pressure disturbance to a region which is either undergoing slight or no strain or to a region which

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is undergoing strain which is opposite in sense. For strains of tectonic origin, the wavelengths of the pressure disturbance are likely on the order of kilometers so that this problem is likely to take place only at very low frequencies. The third source of sensitivity deterioration is vertical flow from or to the water table. Since the water table is particularly insensitive to crustal strain (Rojstaczer, 1987), any hydraulic communication between the water table and the zone of saturated rock monitored can cause significant attenuation of the strain signal. Water wells are typically in communication with rocks which are less than 200 m below the water table, and we might expect that this source would begin to operate at frequencies higher than that of the second source. Because the second and third sources of sensitivity attenuation described above are directly due to large scale dissipation of pore pressure within rock, their effects cannot be mitigated.
The problem of noise in the water level signal has not been quantitatively examined, but the problems of strain sensitivity related to groundwater flow have been studied previously in some detail. Sensitivity attenuation because of fluid flow into and out of the well has been extensively examined (Cooper et al.. 1965; Bodvarrson, 1970; Johnson, 1973; Johnson and Nur, 1978; Gieske, 1986); it will only be peripherally discussed here and is discussed in detail in a related paper (Rojstaczer, 1987). The influence of the water table on attenuation has been examined by Johnson (1973) and Johnson and Nur (1978); these studies theoretically examined this influence assuming that the unsaturated zone above the water table did not significantly influence water table response.
This study examines: the influence of groundwater flow to the water table on the sensitivity of water wells as strain meters; the noise levels of water wells in comparison with other strain meters. The influence of groundwater flow on attenuation of sensitivity is examined by: extending the theoretical work of Johnson (1973) and Johnson and Nur (1978) to include the influence of the unsaturated zone above the water table on atmospheric pressure induced strains; applying these theoretical results to the response of two wells to atmospheric loading in order to examine how groundwater flow influences strain sensitivity as a function of frequency. The comparative

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performance of water wells as strain meters is assessed by calibrating two wells to known strains and examining how noise levels, in terms of strain, increase with time.

OVERVIEW OF WELLS TO BE EXAMINED

Figure 2 shows the water level record for the two wells which are examined here. These wells have been chosen because: they are located in areas which have been the focus of many crustal deformation studies over the past decade; and they exhibit the largest amplitude response to the strain induced by earth tides of any of the wells monitored over the time shown. The hydrograph labeled LKT is the water level record from a well in the Long Valley caldera near Mammoth Lakes, California; the hydrograph labeled TF is the record from a well near the San Andreas fault near Parkfield, California. Both areas are being monitored (with water wells and more conventional strain instruments) in the hope of detecting strains precursory to a tectonic event. Within Long Valley, the possible tectonic event is a major earthquake or a volcanic eruption (Cockerham and Pitt, 1984; Savage and Clark, 1982; Miller, 1985). Near Parkfield, the historic record points to the strong possibility that a magnitude 6 earthquake will take place near Parkfield by 1993 (Bakun and McEvilly, 1984).
If any tectonic event within Long Valley or Parkfield is to be successfully predicted using strain meters it must be preceded by strains which are large enough to be detected with strain instrumentation and have a character which is noticeably different from any background strain that may be present within the region. The water level records are LKT and TF over the second half of 1985 both show a long term decline in water level; this low frequency decline is at least partially due to the relative lack of precipitation in central California over the winter of 1984-1985, but even if this decline was solely due to tectonic strain it would be difficult to utilize such low frequency behavior as a precursor to a major tectonic event. Superimposed upon the trend are deviations on the order of centimeters and which appear nearly at the same time in both wells. These fluctuations of period of several days to several weeks are (as we shall see

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later) in the frequency range in which we can expect to use these water wells as sensitive and relatively noise free strain meters and they are largely due to strains induced by atmospheric loading. Finally, there are high frequency fluctuations (on the order of a cycle/day) which are largely due to atmospheric pressure and earth tide induced strains and are about a centimeter in amplitude.
The above qualitative discussion of the water well records can be brought into somewhat sharper focus by examining the power spectra of the water level records shown in Figure 3. Both wells have considerable power in the semi-diurnal and diurnal frequency band due to earth tides and atmospheric loading. Power levels continue to increase at lower frequencies. This increase in power with decreasing frequency is typical of any continuously monitored strain instrument (Agnew, 1986) although the rate of power increase with decreasing frequency is a function of the instrumentation and background strain rate.
Figure 3 also shows the power spectra of atmospheric pressure at TF and
LKT over the second half of 1985. Comparison of the atmospheric pressure power spectra with the water level power spectra at TF and LKT indicates many similarities. There is substantial power in the semi-diurnal and diurnal band and a strong increase in power with decreasing frequency. At LKT the power spectra of atmospheric pressure and water level have a character which in the frequency band of 0.1 to 0.8 cycles/day is nearly identical. The power spectra of atmospheric pressure and water level at TF are relatively less similar.
In a later section, we quantitatively examine the influence of atmospheric loading on the response of these wells. The response of these water wells to earth tide induced strains while not the primary focus of this study is-used to quantitatively calibrate water level fluctuations in terms of strain.

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INFLUENCE OF THE WATER TABLE ON THE SENSITIVITY OF WATER WELLS TO AREALLY EXTENSIVE STRAINS

As has been previously noted, we cannot expect the response of a water well to strain to be independent of frequency. At low frequencies, the response of a well will be dominated by the influence of the water table and, as a result, will be negligible. At higher frequencies, the influence of the water table will be weaker and we can expect that (until frequencies are so high that flow into the borehole becomes a difficulty) sensitivity will increase with frequency. We assume, for the purposes of this study, that borehole flow (either due to high permeability or to the installation of a packer) does not significantly attenuate strain sensitivity in the frequency range of interest. When the frequency of the imposed strain is high enough to effectively isolate pore pressure from water table influences, we follow hydrologic convention and call the response of the water well the confined response. In this section we examine the response of water wells to periodic crustal strains and stresses as a function of frequency; in the following section we will apply these theoretical results to the observed frequency response of LKT and TF.
In order to examine the influence of the water table on water well
response, we need to determine how pore pressure changes in response to periodic deformation. In a related paper (Rojstaczer, 1987), it is shown that time-dependent, pore pressure response to laterally extensive and uniform deformation depends upon whether the deformation is an imposed areal strain or an imposed surface load. Under conditions of one-dimensional vertical fluid flow, the response of pore pressure, P, at typical well depths to areal strain (sum of the principal horizontal strains), « , imposed by earth tides and broad-scale tectonic deformation is governed by:

D- !2-f! + C fl2r (1)
az 2 at at

where D' is a. hydraulic diffusivity for imposed horizontal strains under conditions of plane stress (Van der Kamp and Gale, 1983; Rojstaczer, 1987):

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C is a measure of the sensitivity of pore pressure to imposed areal dilatation (Van der Kamp and Gale, 1983):

/) - 2aB(l-2ix)

B is a coefficient which relates mean stress to pore pressure under undrained conditions (Skempton, 1954; Rice and Cleary, 1976):

B - 08 - 0) (4)

and a is the fraction of rock strain taken up by the pore space under drained conditions (Biot and Willis, 1957; Nur and Byerlee, 1971):

a - 1 - 0g/0r (5)

It should be noted that p is the fluid density, g is gravity, k is the hydraulic conductivity, ft is the rock matrix compressibility, v is Poisson's ratio, <f> is the porosity, ft- is the fluid compressibility, ft is i. S the rock grain compressibility and /i is the shear modulus.
Under conditions of one-dimensional vertical fluid flow in response to areally extensive surface loading (such as that produced by atmospheric loading), pore pressure response at typical well depths is described by:

_ a 2 P aP . da, , D __ = _ +7 b (o;
az 2 at at

where D is a hydraulic diffusivity under conditions of areally extensive imposed surface stress (Rojstaczer, 1987):

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7 is the near- surf ace response of pore pressure to surface loading under undrained conditions (Rojstaczer, 1987):

5B(l+i/) _________ (8)
+ 2aB(l-2i/)

and a. is the surface load. b
Equations 1 and 6 indicate that pore pressure response to imposed areal strain and surface loading are both described by diffusion equations which contain a source term. The essential difference is that the source term in the case of imposed areal strain has (due to the hydraulic diffusivity, D' , being larger than the hydraulic diffusivity, D, for equal bulk material properties) a weaker influence on time dependent changes in pore pressure. The diffusion equations given above are an accurate description of pore pressure response as long as we can assume that, in the frequency range of interest, there is no lateral pore pressure dissipation; under this condition, the deformation and state of stress at a point are independent of far-field pore pressure (Biot, 1941; Rice and Cleary, 1976). We also assume that material properties are uniform throughout the vertical column of interest. It should be noted that the hydraulic diffusivities D' and D are a strong function of hydraulic conductivity, k, and are also a strong inverse function of rock matrix compressibility.
Figure 4 shows two problems of interest with regard to the influence of
the water table. In the first problem, the rock is subject to a periodic areal strain Acos(wt), the water table is at zero pressure and the well is an accurate gage of pore pressure; this is an idealized description of the response of pore pressure to broad- scale tectonic and earth tide induced strains. The second problem is an idealized description of the water well response to atmospheric loading: the rock is subject to a periodic vertical stress Acos(wt) and areal strain, e« , is one half the cubic strain e (Farrell, 1972); the water table is subject to a pressure -AGcos(wt-0) where G and 0 account for the attenuation and phase shift of the atmospheric pressure due to diffusion of air through the unsaturated zone above the water table; the water in the well is subject to a periodic pressure -Acoswt

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and the relation between water level and pore pressure is (Rojstaczer, 1987):

W = (Acoswt+P)/pg (9)

Since fluid flow is one-dimensional, both problems outlined above are readily solved analytically.

Frequency response of a well to imposed horizontal strain

The response of a water well to imposed periodic areal strain is given
by (Appendix A):

W- [ACexp(-jQ')cos(wt-jQ') - ACcos(wt)]/pg (10)

where A is the amplitude of the dilatation and Q' is a dimensionless frequency:

Q' = z 2 w/2D' (11)

This result is qualitatively similar to the frequency dependent response of water wells to strain given by Johnson (1973) and Johnson and Nur (1978). The major differences between their analogous solution and that given here are due to their approximation of the water table as a spherically shaped boundary. Equation 11 indicates that the most important parameter governing sensitivity as a function of frequency is, z, the depth from the water table. Hydraulic diffusivity is also an important factor with low diffusivity favoring low attenuation at a given frequency. As was previously noted, hydraulic diffusivity is a strong inverse function of rock compressibility and as a result, low attenuation at low frequency is favored for highly compressible rock. Conversely, it is noted that sensitivity due to horizontal loading under confined conditions is strongly favored by low compressibility (Van der Kamp and Gale, 1983; Rojstaczer, 1987). Hence, high sensitivity under confined conditions to strains

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produced by horizontal loading will be accompanied by a relatively rapid attenuation with decreasing frequency.
The gain and phase of the sensitivity to dilatation are plotted as a function of the dimensionless frequency, Q', in Figure 5 under the assumption that the confined sensitivity of the well to rock dilatation is 0.05 cm water level drop per areal nanostrain (ppb). The figure indicates that attenuation and phase shift with decreasing frequency is a gradual process. Attenuation and phase shift begin to significantly deviate from the confined response when the dimensionless frequency decreases to a value of 10. Between a dimensionless frequency of 1 and 10, the strain signal is slightly amplified relative to confined conditions; near complete attenuation of the strain signal takes place for strains with dimensionless frequency 0.001 or less.
We can gain some understanding of how an average well will respond to crustal strain by assuming some values for diffusivity and depth from the water table. Given a well tapping rock with a vertical permeability of 10 millidarcies (the average permeability of non-argillaceous materials inferred for the crust (Brace 1980; Brace, 1984), a compressibility of 10" 11 cm2 /dyne, a porosity of 0.10, and a well depth (relative to the water table) of 100 m, the dimensionless frequency will have a value greater than 10 for strains with frequency greater than 20 cycles/day. This result indicates that, for reasonable material properties and geometries, the water table can have a strong influence on strain sensitivity in the frequency range of practical interest.

Frequency response of a well to atmospheric loading

The response of a well to atmospheric loading is considerably different than that given above. These differences due to the stress placed upon the earth's surface and the diffusion of the load through the air phase of the unsaturated zone (Weeks, 1979) Water level response to periodic fluctuations in atmospheric loading is given by (Appendix B):

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W -[(-M+-y)Aexp(-jQ)cos(wt-jQ) - (-y-l)Acos(wt) - NAexp(-jQ)sin(o>t-jQ)]/pg (12)

where M and N are:

M = 2cosh(Jjncosdir)_
cosh (2j R) +cos (2j R)
(13)
N = 2sinh(JiOsln(jR)_
cosh(2jR)+cos(2jR)

Q and R are dimensionless frequencies referenced to fluid diffusivity, D, and air diffusivity, D , respectively.:

3.

Q - z 2 w/2D

R - L2 w/2D (14)

3.

and L is the depth from the earth's surface to the water table. The gain (barometric efficiency, BE) and phase of the response of a water well to atmospheric pressure fluctuations is shown in Figure 6 for a well with a surface loading efficiency, -y, of 0.5.
The response is plotted as a function of two dimensionless parameters: dimensionless frequency, Q, and the ratio of dimensionless frequencies R and Q. The dimensionless ratio, R/Q, is a measure of the time taken for atmospheric pressure changes to reach the water table versus the time taken for water table effects to significantly influence water well response.
For values of R/Q less than 0.0001, the water table is fully influenced
by atmospheric pressure changes at dimensionless frequency, Q, and the response of the water well to atmospheric loading is functionally identical (except for a phase shift of 180°) to the response to strains which are applied directly to the solid phase alone. Under these conditions, the attenuation of sensitivity to tectonic or earth tide induced strains can be inferred from the air pressure response once an allowance is made for the slightly different hydraulic diffusivity which governs fluid flow. For values of R/Q greater than 0.1, there is significant attenuation and phase

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shift of the atmospheric pressure signal at the water table; under these conditions we can expect that, over a frequency band whose width and location is a strong function of R/Q, the water well response will be significantly amplified relative to the confined response. When R/Q is greater than 10, we can expect a water well to respond significantly to atmospheric loading in a frequency band where response to tectonic or earth tide induced strains are strongly attenuated. Hence, the response of a water well to atmospheric loading, when R/Q is significantly greater than 0, is not indicative of how the sensitivity to tidal or tectonic strains attenuates with frequency.
Although it is not possible, based upon the response to air pressure,
to directly infer how the sensitivity of water wells will attenuate in response to imposed strain when R/Q is 0.1 or greater, the air pressure response gives us an indirect means to determine how wells respond to tectonic strain. The material and fluid flow properties of the rock and/or sediment which control well response can be determined by fitting the atmospheric pressure response of the well to the theoretical response (equation 12 or B12). These elastic and fluid flow properties can then be used in conjunction with the theoretical response to imposed areal strain (equation 10 or A7) to infer how sensitivity to tectonic strain attenuates with frequency. In the following section, we examine the atmospheric pressure response of LKT and TF to see if any air diffusion effects are present and use the atmospheric pressure response to infer how these wells respond to tectonic strain.

BEHAVIOR OF WATER WELLS AS STRAIN METERS IN PRACTICE

The previous section focused on the theoretical behavior of water wells
in response to strain. In theory, water wells may suffer from relatively rapid attenuation of sensitivity in a frequency range which is of practical interest for purposes of deformation monitoring. In this section we examine the behavior of two wells, LKT and TF, to earth tide and atmospheric pressure induced strains to determine whether the previous theoretical results have any merit in practice. We also try to examine the overall

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noise levels of these wells as a function of frequency. The wells, as previously noted, exhibited the highest sensitivity to earth tides of all the wells monitored for tectonic strain in the Long Valley caldera and near Parkfield, California in the second half of 1985 and it might be expected that the sensitivity and noise levels of these wells are close to the best response we can expect from wells which are no more than 300 m deep.
The depth and near-hole lateral permeabilities of these two wells are shown in Table 1. Both wells are isolated from the near surface to a depth in excess of 100 m. The depth to water is about 20 m at both LKT and TF; if near-hydrostatic conditions prevail at these wells, then this depth indicates the depth to the water table at both sites. The lateral permeabilities of the rock in direct hydraulic communication with these wells were inferred from a slug test (Kipp, 1985) and pumpage data from LKT and TF respectively. These permeabilities are high in relation to the range of permeabilities which have been measured or inferred for the crust (Brace, 1980) indicating that hydraulic communication between the well and rock is relatively good. The borehole diameter is roughly 15cm at both wells.
At each site, atmospheric pressure and water level were measured with silicon strain bridge transducers 4 times per hour. The water level and atmospheric pressure records over the time interval shown in Figure 2 contain 2% and 5% gaps at LKT and TF respectively; these gaps range from 1/2 hour to a few days in length. The method by which these gaps in the time series were filled is described in Appendix C. The tidal areal strain time series at each site was determined from the theoretical tidal potential with no corrections made for ocean loading, topographic or geologic effects. The results of Beaumont and Berger (1975) and Berger and Beaumont (1976) suggest that the amplitude of actual tidal dilatation agrees with those determined from the homogeneous earth tide with an error of about ±50%.
The relation of water level to crustal strains induced by atmospheric pressure and earth tides was determined by cross-spectral estimation (Appendix D). Table 2 indicates the response of these wells to the M2 and Oj earth tides in terms of their dilatational efficiency (water level drop in centimeters per areal nanostrain) and phase. Their response to the M2 tide in terms of dilatational efficiency is identical; for both sites the

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dilatational efficiency with regard to the M2 component of the tidal potential is about 0.034 cm water level drop per areal nanostrain. The dilatational efficiency of the 0 1 tide is slightly lower, particularly at TF. At TF, this slightly greater difference in amplitude may be partly due to water table influences which (as is shown below) significantly attenuate strain sensitivity in the frequency band of interest (0.025 to 2.5 cycles/day). Phase differences at LKT and TF between the 0 1 and M2 tidal constituents are slight. The constituents at LKT are both roughly -10° out of phase with the response that would be expected if the phase of the tidal dilatation could be precisely determined from the theoretical tidal potential; at TF, the phase of the 0- and M9 tidal constituents are both roughly +10° out of phase with the theoretical response. The phase shift of both constituents as well as the small difference in dilatational efficiency between the ML and 0- constituents at LKT are likely due to local inhomogeneities and/or ocean loading effects. At TF, the phase shift of both constituents is likely due to a combination of local inhomogeneities, ocean loading effects and water table influences.
The response of LKT to atmospheric loading is shown in Figure 7. The phase is generally flat and near 0° out to a frequency of 0.02 cycles/day. The admittance or barometric efficiency begins to show some attenuation below a frequency of 0.05 cycles/day. This attenuation may not reflect any water table influence; rather it may be the result of some error in the estimate of barometric efficiency at low frequencies. Bendat and Piersol (1986) give error estimates for the admittance and phase determined from cross-spectral estimation. For this study the 95% confidence interval for admittance or barometric efficiency, BE, is given by:

BE 95 = BE ± .87j(l-r 2) (15)

where F is the coherence. Figure 8 shows the coherence squared, as a function of frequency for LKT. At frequencies less than 0.05 cycles/day, the coherence squared is substantially less than 1 and the error bounds on barometric efficiency are greater than ±.38. A flat barometric efficiency of .48 (the mean value of the barometric efficiency for frequencies greater

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than 0.05 cycles/day) over the entire frequency range examined is thus not inconsistent with the admittance values and is consistent with the general frequency insensitivity of the phase. Alternatively, the attenuation which begins to occur at frequencies less than 0.05 cycles/day is a real physical phenomenon, presumably due to the influence of the water table.
The fit to the data, based on the assumption that there is some water table influence in the observed frequency band is also shown in Figure 7. The best fit to the data is achieved with a confined barometric efficiency of .45 (7-.55), a value for dimensionless frequency Q of 30.Ow, and a value for dimensionless frequency R of 5.9w where frequency w is in units of cycles/day. The value for the confined barometric efficiency given by this fit combined with the dilatational efficiency estimate for the M~ tide shown in Table 2 allow us to calculate (Rojstaczer, 1987) a compressibility for the rock in communication with the well. Assuming a Poisson's ratio of .25, the drained rock compressibility is 1.3 x 10 1X cm2 /dyne. Given this compressibility and assuming that the depth to water is indicative of the depth to the water table, it is also possible to estimate the vertical permeability of the zone between the water table and the uppermost depth that the well is in hydraulic communication with saturated rock. This permeability is 0.8 millidarcies which is considerably below the lateral permeability given in Table 1; the difference indicates that either moderately low permeability layers exist above the monitored zone or that there is considerable anisotropy in the permeability of the rock and sediment above the zone monitored. It should be noted that this inferred permeability is a maximum permeability based upon the assumption that the attenuation indicated at low frequencies is a real phenomenon. If significant attenuation is not present in the entire frequency band analyzed, vertical permeability would be lower.
The elastic and fluid flow properties determined from the response of
LKT to atmospheric loading are used to infer how this well responds to crustal strains (Figure 9; note that we assume that the attenuated response to atmospheric loading reflects a real physical phenomenon). In terms of attenuation of sensitivity, it is not considerably different from the atmospheric pressure response. This correspondence indicates that the water

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table is largely in phase with the pressure disturbance. The inferred sensitivity remains high when frequencies are less than 0.001 cycles/day. Unfortunately, it will be shown below that while sensitivity remains high for low frequency response, the noise level at very low frequencies is also high.
The response of TF to atmospheric loading is shown in Figure 10. This response has a signature which indicates that water table influences are present throughout the observed frequency band. The relative lack of ambiguity in this signal is evident in the coherence squared for the transfer function (Figure 8) which is near 1 for frequencies in excess of 0.06 cycles/day. The response to atmospheric pressure has a maximum at about 0.5 cycles/day. It might be suspected that decreasing sensitivity with increasing frequency above the frequency of 0.5 cycles/day would be at least partially due to limited groundwater flow into and out of the borehole; this source of attenuation is unlikely because it is inconsistent with the near flat dilatational sensitivity for the semi-diurnal and diurnal tidal constituents shown in Table 2 and is also inconsistent with the phase of the atmospheric pressure response. Rather, this decreasing sensitivity at higher frequencies is likely due to air diffusion effects which amplify the atmospheric pressure response in the frequency band of 0.2 to 2 cycles/day. As Figure 6 indicates, such a response would be theoretically possible if the dimensionless numbers, R and Q, were of the same magnitude. The best fit to the data is given with a confined barometric efficiency of 0.37 (7=.63) and a value for both Q and R of 2.2co where frequency is in terms of cycles/day. This value of confined barometric efficiency and the value of dilatational efficiency given in Table 2 for the M_ tidal constituent indicate that the drained rock compressibility for the rock in direct communication with the well bore is 1.5 x 10 ll cm2 /dyne. The vertical permeability above the well is estimated to be 10 millidarcies, a value which is only slightly less than the lateral permeability inferred from pumping data, indicating that the material above and within the zone monitored is largely homogeneous and isotropic with respect to permeability. The inferred response to tectonic strain is shown in Figure 9. In terms of attenuation of sensitivity, it is considerably different from the

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atmospheric pressure response. The inferred response to strain approaches confined conditions at a frequency of 2.5 cycles/day but is never fully observed in the frequency band analyzed. Below a frequency of 0.001 cycles/day where strain sensitivity is about 1/10 the inferred confined response, strain sensitivity asymptotically approaches zero.
The raw and filtered strain spectra for LKT and TF are shown in Figure
11. These strain spectra were derived by: 1) obtaining power spectral densities for the water level records in terms of cm2 per cycle/day; 2) converting these spectra into units of strain2 /Hz through the use of the M2 dilatational efficiencies given in Table 2; 3) adjusting the strain spectra, when necessary, to account for any unambiguous frequency dependent changes in dilatational efficiency inferred from the response to atmospheric loading; 4) normalizing the spectra relative to 1 strain2 /Hz and converting to a decibel scale. The spectral estimates TF were adjusted upwards to account for the observed decreasing sensitivity with period by essentially adding 6 db per decade to the slopes of the strain spectra for frequencies less than 1 cycle/day; since the observed attenuation at LKT is slight and ambiguous in the frequency band examined here, no adjustments in the strain spectra were made. The raw strain spectra at LKT and TF are very similar and increase at a rate of roughly 25 db per decade in frequency, a rate that is slightly high compared to the best dilatometers (Johnston et al.. 1986) but is comparable to the behavior of some laser strain meters (Beavan and Goulty, 1977). It should be noted that some of this increase in power with decreasing frequency at LKT is due likely to tectonic strain; extensional strain rates in the Long Valley caldera were about 0.5 ppm/year during the second half of 1985 (Langbein et al.. 1987a; Langbein et al.. 1987b).
Considerable reduction in noise level can be achieved by removing the effects of barometric pressure on water level response. These effects were removed in two ways. In the first method the fit of the observed frequency response shown in Figures 7 and 10 was assumed to be the transfer function between water level and atmospheric pressure. This transfer function was then multiplied with the Fourier transform of the atmospheric pressure record and the resultant frequency response was then inverted into the time

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domain and subtracted from the water level time series. In the second method a single coefficient between the water level record and the atmospheric pressure record was found by linear regression of the two time series (the coefficient was .48 for LKT and .35 for TF); the atmospheric pressure record was then simply multiplied by this coefficient and the resultant time series was then subtracted from the water level record.
At LKT, there is little difference between the strain spectra for the time series obtained with regression and the frequency dependent transfer function in the frequency range of interest. The lack of difference between the two strain spectra is due to the relatively flat response to air pressure in the frequency range of interest; as a result, it is operationally similar to the filter which assumes a simple linear relationship between water level and atmospheric pressure.
At TF, there is substantial difference between the strain spectra determined with the regression coefficient and the frequency dependent transfer function because there is considerable attenuation and phase shift of the atmospheric pressure signal.
Both filtered records at LKT and the record filtered with a frequency dependent transfer function at TF yield strain spectra whose slope increase at a rate of roughly 20 db per decade in frequency in the frequency band of 0.08 to 2.5 cycles per day. This is comparable to the rate of increase seen in the raw strain spectra of high quality strain instruments (Agnew, 1986; Beavan and Goulty, 1977; Johnston et al., 1986). Presumably, one could also remove the effects of atmospheric pressure from other strain instruments as well to decrease the rate at which noise levels increase. For instruments which measure horizontal strain only, however, the reduction in noise rate achieved by removing atmospheric loading effects would be slightly less than that for water wells since the sum of the principal horizontal strains induced near the earth's surface by atmospheric loading is approximately 1/2 the cubic strain (Farrell, 1972). The similarity in rate of noise increase between water wells and other strain instruments suggests that either the effects of precipitation on the water level record are slight in this frequency band or are not significantly greater than other strain meters (Takemoto, 1983). Below 0.08 cycles/day, the slope of the corrected strain

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spectra increase significantly at both sites to about 35 db per decade. This slope increase may be due to the influence of precipitation, but the limited amount of data present for frequencies less than 0.1 cycles/day in these strain spectra make any analysis of low frequency behavior ambiguous. Following the approach of Agnew (1987), we compare the strain response shown in Figure 11 with standard electronic distance measurement response by transforming standard error estimates for distance measurements into strain spectra. When the strain spectrum of the water well exceeds the strain spectrum of the distance measurement, the water well is less accurate; when the strain spectrum of the well is less then the strain spectrum of the distance measurement, the water well is more accurate. The most accurate electronic distance measurement currently available is obtained with a two color laser device (Linker et al.. 1986). The standard error for a two color distance measurement taken over a baseline of 10km is 1.2 x 10 7. If 10 km length distance measurements were made to determine areal dilatation with a two color geodimeter once a day, the resultant strain spectrum would be (in the absence of a tectonic signal) a flat -83 db, relative to 1 strain2 /Hz. The raw and corrected strain spectra for both wells are well below this noise level until frequencies are less than 0.04 cycles/day indicating that these wells are quite capable of outperforming existing distance measurement capability over the time span of days to weeks.

CONCLUSIONS

The results given here suggest that theoretical models which describe
the response of water wells to air pressure and tectonic strain can be used, in conjunction with cross-spectral estimation, to yield some valuable information on the influence the water table has on strain sensitivity. When the water table is weakly isolated from air pressure changes, the attenuation of the air pressure response of water wells is qualitatively similar to the attenuation of the earth tide and tectonic strain response. Well sensitivity in both cases gradually attenuates with decreasing frequency due to the presence of the water table and the frequency at which significant attenuation begins to take place is a strong function of well

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geometry and rock material properties. For wells open to rock in excess of 100 m below the water table, attenuation of sensitivity will be slight at periods of days to weeks if the rock above the open interval has a vertical permeability of 0.1 md or less and a compressibility of 1 x 10 11 cm2 /dyne. The wells examined here are most useful as strain meters over a limited frequency band. Both exceed the accuracy of standard distance measurement techniques for strains with periods of days to weeks; at longer periods, electronic distance measurements can have superior performance. At TF, increasing noise with increasing period appears to be the result of decreasing strain sensitivity in the presence of a 15 to 20 db per decade increase in the water level power spectrum. At LKT, the increasing noise appears to be largely due simply to a 20 to 25 db per decade increase in the water level spectrum; since some portion of the drift at LKT is likely due to tectonic strain, the noise levels noted for LKT may be underestimated. The noise levels of these wells, in the frequency band of 0.08 to 2.5 cycles/day are not substantially different from the noise levels of other continuous strain meters; this correspondence indicates that if precipitation influences these wells in this frequency band, it does so at a level which is not significantly different than its influence on other strain instruments. If tectonic events are preceded by strains on the order of 10 8 which take place over periods of less than a month, they can be detectable in water wells which are sensitive to strain.

APPENDIX A: SOLUTION TO THE RESPONSE OF A WELL TO A PERIODIC DILATATION WHICH DOES NOT DIRECTLY PRESSURIZE THE WELL OR WATER TABLE

The solution to the response of a well to periodic dilatation is obtained by solving for the pore pressure response to periodic dilatation. We assume that the well is an accurate gage of the average pore pressure of the saturated rock with which it is in communication and assume that the open interval of the well is very small relative to the change in pore pressure with depth. Pore pressure response to a periodic dilatation, Acoswt, is governed by the equation:

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3 2 P 3P ' = _ - CAwsinut (Al)

Equation Al was obtained from equation 1 by substituting Acoswt for e- . This equation must be solved subject to the boundary conditions:

P(0,t)= 0

P(«,t)= -ACcoswt (A2)

where z=0 is taken to be the water table. No initial condition is imposed because we seek the periodic steady-state solution. This problem is easily solved by employing complex notation. Taking P to be complex:

P(z,t)= F(z)exp(iwt) (A3)

and substituting in equation Al we obtain:

F" = r iwF + ~ icoCA ... . (A4)

F(0) = 0
F(«) = -AC

where, ', implies differentiation and all exponential terms have been divided out. Equation A4 is a second order inhomogeneous ordinary differential equation. Its particular solution, F , is:

F - -AC (A5)
P

Its homogeneous solution, F, , is:

R = ACexp(-(i+l)jQ') (A6)

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where Q' is defined in equation 11. Summing equations A5 and A6, the solution in terms of gain and phase is:

Gain = |P/AC| = J(J 2 +K2) (A7)

Phase = Tan-1 (K/J)

where J and K are:

J=exp(-jQ')cos(jQ') - 1
_ (A8) K=exp(-jQ')sin(jQ')

Since the well is assumed to be an accurate gauge of pore pressure in the frequency band of interest, water level changes are related to pore pressure changes by W=P/pg and the solution in terms of water level change per change in strain can be obtained from equation A7 by multiplying the gain by C/pg. The solution in the real domain is given in equation 10.

APPENDIX B: SOLUTION TO THE RESPONSE OF A WELL TO PERIODIC FLUCTUATIONS IN ATMOSPHERIC PRESSURE

The solution for the response of a well to periodic fluctuations in atmospheric pressure is obtained, as in Appendix A, by solving for the pore pressure response. In Appendix A, it was assumed that the water table (z=0) was always at zero pressure; this is not the case for periodic fluctuations in atmospheric pressure and before pressure at depth can be known the water table pressure must be determined. The water table response to periodic fluctuations in atmospheric pressure is determined through the use of a diffusion equation for flow of air through unsaturated porous materials (Buckingham, 1904; Weeks, 1979):

n f*i\ D __a= _a (Bl)
a at

subject to the following boundary conditions:

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P (-T,t) - Acoswt
a (B2)
P (T,t) - Acoswt

cL

where P is the air pressure. The boundary -T is taken to be the earth's surface; the zone from depth 0 to depth T is simply an artifice to assure that at the water table, z-0, there is no air flux. As in Appendix A, we seek the periodic steady state solution:

P (z,t) - F(z)exp(iwt) (B3)

3.

Substitution of B3 into equations Bl and B2 yields the following second order homogeneous differential equation with transformed boundary conditions:

F" - iwF
D (B4)

P (-T,t) - P (T,t) - A

cl cL

The solution for the air pressure at the water table is obtained by solving equation B4 at z=0:

P - MAcos(wt) + NAsin(wt) (B5) a

where M and N are given in equation 13. The pore pressure at the water

table is -P . a
With the pressure at the water table known, the solution to pore pressure response to a periodic atmospheric pressure fluctuation, Acoswt, is obtained from the following modified version of equation 6:

D - - -yAwsinwt (B6)
dz 2 at

The appropriate boundary conditions are:

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P(0,t)- -MAcos(wt) - NAsin(wt)

P(«,t)- -A-ycosut (B7)

We again take P to be complex:

P(z,t)- F(z)exp(iwt) (B8)

and substituting into equation B6 we obtain:

(B9)

F(0) - (-M+iN)
F(oo) - -A7

where, ', implies differentiation and all exponential terms have been divided out. Equation B9 is a second order inhomogeneous ordinary differential equation. Its particular solution, F , is:

Fp - -A7 (BIO)

Its homogeneous solution, F, , is:

Fh "

The sum of equations BIO and fill multiplied by exp(iwt) yield the solution to pore pressure response to periodic changes in atmospheric pressure. The response in terms of water level within the well can be obtained (as in equation 9) by adding Acosut to the real part of the solution and dividing by pg. The solution for water well pressure, P , in terms of phase and gain is given by:

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Gain = |P /A| - J(U2 +V2)

Phase - Tan-^V/U)

(B12)

where U and V are:

U=(-M+-y)exp(-jQ)cos(jQ) + Nexp(-jQ)sin(Jbj - (7-!)

V=(-M+-y)exp(-jQ)sin(jQ) - Nexp(-jQ)cos(jQ)

(B13)

The solution in the real domain in terms of water level is given in equation 12.

APPENDIX C: METHOD BY WHICH GAPS WERE FILLED IN THE WATER LEVEL AND ATMOSPHERIC PRESSURE RECORD

Gaps in the water level and atmospheric pressure record over the time period examined were filled by an iterative process. The gaps were originally filled by linear interpolation. A finite impulse response high pass filter (Otnes and Enochson, 1978) with a cutoff frequency of 10 8 Hz was then applied to the water level and atmospheric pressure data to remove any long term trends. The autocovariance with a maximum length of 40 days was then calculated for each time series. Gaps were then filled with a symmetric linear filter of length 120 with weights determined from the structure of the autocovariance and the distance between the interpolated point and the nearest data point. The spacing between each value used in the filter was eight data points (two hours) and weights were calculated by solving the following system of linear equations:

C lf2

C 2! 1

c eo t i
1 1

C l t 60 I
C 2 t 60 I
1

a 2 1
1 0

wi

W2

. .

W60
_ M _

=

C J t'
C 2 t'

. . C 60 t ' _ 1

(Cl)

where a 2 is the variance, n is the mean, Citj is the autocovariance between the ith data value and the jth data value used in the interpolation, w.is the weight of the ith data point in the filter and Cit' denotes the

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autocovariance for the length between the interpolated point and the ith nearest data value. This linear system of equations provides an estimate of the interpolated value for which the estimation variance is at a minimum (Journel and Huijbregts, 1978). The estimation variance is a 2 is defined as:

a 2 = a 2 + /* - Zw.Cit' (C2)

6 -L.

Once the gaps were interpolated using the system of equations given in equation Cl, the residual time series (the original time series with linearly interpolated gaps minus the high pass filtered time series with linearly interpolated gaps) was added to the new time series to preserve the long term trend.

APPENDIX D: METHOD BY WHICH THE TRANSFER FUNCTIONS OF WATER LEVEL TO EARTH TIDES AND ATMOSPHERIC PRESSURE WERE DETERMINED

The relation of water level to earth tides and atmospheric pressure at
the wells in the frequency domain was obtained by solving the following complex system of equations for every frequency (Bendat and Piersol, 1986):

BB BT HB

TB TT HT (Dl)

where: BB and TT denote the power spectra of the atmospheric pressure and earth tides respectively; BT denotes the cross spectrum between atmospheric pressure and earth tides; TB denotes the complex conjugate of the cross spectrum between atmospheric pressure and earth tides; BW and TW denote the cross spectra between atmospheric pressure and water level and earth tides and water level, respectively; and HB and HT denote the transfer function between water level and atmospheric pressure and water level and earth tides respectively. The power spectra and cross spectra were obtained for the time series of interest by using the Blackman-Tukey procedure (Bendat and Piersol, 1986) after removing the mean and long term trend from the time series. This procedure is computationally inefficient because it requires that the autocorrelation or cross-correlation function be calculated in the

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real domain in order to obtain spectral quantities; it was used because the frequencies of interest are low relative to the length of the data set. The spectral estimates obtained from the Blackman-Tukey procedure are smoothed using a Hanning window (Otnes and Enochson, 1978).
At TF, the long term trend was removed with a high pass finite impulse response filter (Otnes and Enochson, 1978) with a cutoff frequency of 10" 8 Hz. Spectral estimates were then obtained using a using a maximum correlation length of 40 days. At LKT, the time series for water level, atmospheric pressure and earth tides was lengthened to include the first half of 1986 and the data was decimated to 2 samples per hour; the time series was lengthened to see whether any attenuation of sensitivity could be unambiquously identified at low frequencies. The long term trend was removed with a high pass finite impulse response filter with a cutoff frequency of 5 x 10 9 Hz. Spectral estimates were obtained using a maximum correlation length of 80 days.
The barometric and dilatational efficiency as a function of frequency
are simply the gains of the transfer functions HB and HT respectively. The phase relations given for LKT and TF are the phase of the transfer functions. The multiple coherence squared, F 2 , of the transfer functions as a function of frequency is obtained from the following spectral estimates (Otnes and Enochson, 1978):

F 2 = (WB x TT - WT x TB)BW + (WT x BB - WB x BT)TW («.
(TT x BB - TB x BT)WW ('

where WB and WT denote the complex conjugates of BW and TW respectively and WW is the water level power spectrum. This coherence squared is a measure of the ability of a linear relationship of water level to atmospheric pressure and earth tides to account for the water level power.

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Table 1. Description of wells examined. Local permeability refers to the permeability near the borehole.

WELL ID LOCATION OPEN INTERVAL Cm) LOCAL PERMEABILITY

LKT Long Valley,CA 152-296

TF Parkfield, CA 152-177

2 x 10 4 millidarcies

2 x 10 1 millidarcies

Table 2. Tidal response of wells examined. Gain expressed in terms of
centimeters of water level drop per areal nanostrain. Phase expressed in degrees.

WELL ID Q!. GAIN 0 1 PHASE M2 GAIN M2 PHASE

LKT 0.0314 -193.8 0.0337 -189.4

TF 0.0293 -171.1 0.0337 -171.8

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Free surface

Time dependent
water level change

Unsaturated zone

Water table

Attenuation due to
water table effects

"X5Jtt« Attenuation due to
2 flow into region
undergoing extension

Attenuation due to
bore hole flow effects"

Rock undergoing extension

Figure 1 Cross-section of well responding to deformation and principal sources of attenuation of well response.

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Q
Q_

O *
O
O.OOli
0.0001
0.025

O
QL

O *
O
O.OOli
0.0001
0.025

1 2.5
CPD

1 2.5
CPD

Figure 3 Power spectral densities for water level (a) and atmos-pheric pressure (b) at LKT; power spectral densities for
water level (c) and atmospheric pressure (d) at TF. The abbreviation 'CPD' in the figures denotes cycles per day

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Free surface

Water table

P=0
P = 0

= A cos(a»t)

i i

P=-A cos(wt)
Free surface

Water table P = -AG cosM-0)

CTV = A cos (<ot)

Figure 4 a) Idealized model of response of well to areally exten-sive periodic strain; b) Idealized model of response of
well to areally extensive periodic atmospheric pressure changes.

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a

>_ 0.8-
O

+ DATA

MODEL AIR

0.00025 0.001 0.01

CYCLES/DAY

1 2.5

Figure 10 Response of TF to atmospheric pressure in terms of barometric efficiency (a) and phase (b). Fit to data is
solid line denoted as 'MODEL AIR'.

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go

a
o

-70

-80-

-140
0.025

CYCLES/DAY

Figure 11 Power spectral densities at LKT (a) and TF (b) normalized to 1 strain2 /Hz. 'RAW curve is water level record without atmospheric pressure effects removed. 'FILTERED' curve is water level record with atmospheric pressure effects removed with a frequency dependent
transfer function. 'SIMFILT' curve is water level record with atmospheric pressure effects removed with a filter of length one.

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GROUNDWATER-LEVEL MONITORING FOR PREDICTION OF EARTHQUAKES

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ABSTRACT

Since 1976 groundwater-levels and temperature of thermal waters have been monitoring in 100 wells distributed mostly in the southern Kanto and Tokai district, Japan for prediction of earthquakes. Many good examples of precursors were recognized in the following destructive earthquakes: the 1978 West Off Izu-Oshima earthquake(M 7.0), the 1978 Off Miyagi Earthquake(M 7.4), the 1980 East Off Izu Peninsula Earthquake(M 6.7). In the 1978 West Off Izu-Oshima Earthquake, there was a good conformity in precursory behaviors of our water-levels and records of the Japan Meteorological Agency(JMA) bore hole volume strainmeters located in Ajiro and Irozaki, Izu peninsula. In the other cases, precur-sory behaviors were observed by only water-levels and temperature monitoring, but not the bore hole volume strainmeters. The water-level anomalies appeared before M 5 - 6 earthquakes were 1 to 50 cm, which may correspond to 10" 8 to 5 x 10~ 7 in the volume strain-meters. The signals of water-level anomalies were almost equal to the fluctuation of the earth tide and barometric pressure change, and thus might be eliminated as noise or unable to be detected. A relation between magnitude of earthquake and epicentral dis-tance of the farthest groundwater-level anomaly for M 5 earthquakes is 50 km, for M 6 earthquakes is 100 km, for M 7 earthquakes is 200 km, and for M 8 earthquake is estimated to be 400 km. A relation between magnitude of earthquake and precursor time of water-level anomaly for M 5 is 10 days, for M 6 is 21 days, for M 7 is 45 days, and for M 8 is 98 days. When magnitude of earthquake is below 6.5, no intermediate-term precursor{years

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to month) of water-level can be seen,but water-level monitoring is the most promising method to obtain short-term precursors. The intermediate-term precursors in water-levels will appear for the earthquakes magnitude above 6,5.

INTRODUCTION

Since 1976, the Catfish Club, organized with 100 ordinary people, has been monitoring the groundwater-level and temperature of thermal water to observe precursors of destructive earth-quakes. It was one of our motivations to organize the Catfish Club that several warnings of large earthquakes M 7-8 would be expected in the southern Kanto to the Tokai districts in near fu-ture by the authorities. The warnings were based on the measure-ment of land-deformation and on the principle of periodic occur-rence of big earthquakes in the districts during the historic times. They said in addition that the successful prediction could not achieved until early 21st century. Therefore, the local people were seriously threatened by the warnings. We were also encouraged by the fact in China several big earthquakes were successfully predicted through the cooperation of amateur ob-servers and scientists in monitoring of water-level, earth cur-rent, terrestrial magnetism as well as animal behaviors. Thanks to the members' volunteer activity, several good examples of precursors were recognized before destructive earthquakes such as the 1978 West, Off Izu-Oshima Earthquake (M 7.0), the 1978 Off Miyagi Earthquake (M 7.4), the 1980 East Off Izu Peninsula Earthquake(M 6.7), the 1984 Tanzawa Earthquake (M 5.2) and the 1984 West Nagano Prefecture Earthquake (M 6.8). As a result of our 10 years observation, groundwater monitoring has been found to be one of the most significant methods for intermediate- to short-term earthquake predictions.

MONITORING METHOD

In order to keep the accuracy of measurement, a digital water gauge designed by Oki fastened on a steel rig 1.5 m high is installed in each monitoring well. The water gauge is non-

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electric one composed of a tachometer coupled with gears and a pulley strung to a globular float. Its sensitivity is 1 mm. Tem-perature of thermal water is measured with a digital thermometer in sensitivity of 0.1°C. The water-level or temperature is measured once or twice a day together with barometric pressure and precipitation. The observed data are filled in a formulated post-card and mailed once a week, putting additional comment to the headquarters of the Catfish Club, Hot Springs Research In-stitute. The received data are compiled in a hydrograph added felt earthquakes occurred nearby every month and sent back to each observer. Every four months the observed results are sum-marized on a Black and White Score Table in the Catfish Letters, a special issue of the Bulletin of the Hot Springs Research In-stitute for distribution. In case of sudden appearance of large anomalies, the fact is telephoned to the headquarters. After careful check, if necessary, the information is telephoned to each observer for careful observation. Since 1982 a precise thermometer with sensitivity of 0.01°C has been installed in Kurokawa's thermal well at Usami by Mogi, Earthquake Research In-stitute of Tokyo University.
Most of our wells once used for domestic drinking' water are shallow ones, ranging from 5 to 20 m deep. Barometric pressure change is weakly reflected in wells deeper than 10 m. The effect of the earth tide is seen in wells deeper than 100 m; in such cases the water-level check once or twice a day will be insuffi-cient to figure the real behavior of water-level. For deep wells a continuous water-level recorder equipped with a digital water-level gauge is required, but we do not have it yet.
Lamar et al.(1982) and Merifield and Lamar{1983) recognized
the appearance of large spike of water-level change 45 cm high 3 days prior to the 1980 M 5.5 earthquake occurred in the San Jacinto fault zone, south California by the Stevens Type F water-level recorder. In our monitoring system, however, such rapid changing T>7ater-l evel anomalies are difficult to observe.
The most influential factor to the water-level change in shallow wells is precipitation. Particularly, two rainy seasons

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in Japan, early summer and fall are very difficult time to get smooth hydrograph due to disturbance by precipitation. Several deep wells more than 200 m deep protected with good casing and cementing are strongly required as the key wells for water-level monitoring.

CASE STUDY

1978 West Off Izu-Oshima Earthquake(M 7.0)
The Geographical Survey Institute demonstrated in 1977 that there was unusual ground uplift by 15 cm at maximum in the I to area, northern Izu peninsula from 1967 to 1976 (Figure 1). This rapid areal uplift was accepted as a precursor of large earthquakes in the northern Izu area.
This quake occurred on January 14, 1978 under the sea-floor
3 km deep between Oshima Island and the Izu peninsula, and killed 25 persons, collapsed 94 houses and caused several landslides on steep slopes in the middle portion of the Izu peninsula. The epicenter was about 30 km southeast from the cen-ter of uplifted area. Figure 2 shows distribution of the epicen-ter and water-level monitoring wells of the Catfish Club; solid circles indicate precursors observed.
Figure 3 is a summary of groundwater anomalies observed before the main shock. In the area within 50 km from the epicen-ter, distinct precursors appeared in temperature of thermal water and groundwater-1eve1 more than a month before the main shock(Figure 4).
Curve A in Figure 4 is temperature of thermal water naturally discharging from a well 620 m deep at Usami, a fishing port located 36 km north of the epicenter. Y.Kurokawa, owner of the well, has long studied the relation between temperature of this thermal water and earthquakes. He recognized that the sud-den rising of temperature started on December 30 after its gradual decreasing was a precursor of earthquake nearby(Kurokawa 1978). The temperature fluctuation within 0.1°C, which showed good correlation with the fluctuation of barometric pressure meant his temperature measurement was very precise. According to

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his discharge test on the well, 0.1°C increase corresponds to 7.7 cm increase in water-head. The temperature increase in the final phase of the precursor by 0.7°C was the increase of water-head by 54 cm before the quake. There was pronounced coseismic event in the thermal well. Immediately after the main quake, the water temperature dropped by 1.0 *C. From January 15, temperature rose above 60° to finally 61°C with increasing of discharge. Curve B is the water-level fluctuation observed by the staff of Fire Sta-tion of Atami City(1978) in a shallow well 6 m deep, located 50 m from the sea and 40 km north of the epicenter. Beginning on December 9 in 1977, a rise in water-level unrelated to rainfall dr barometric pressure appeared. From December 21 the water-level showed a sawtooth fluctuation with 5 to 15 cm rise each day, and then after 36 days since the beginning the 1978 West Off Izu-Oshima Earthquake occurred. This anomalous fluctuation fur-ther continued after the quake and gradually vanished by the end of February. The installation of a water-level recorder in this well soon after the shock revealed that the high tidal fluctua-tion of seawater was reflected on the high groundwater-level. No other hot springs or wells were present in the area, the change of the water-level could not be influenced by pumping. The water-level anomaly is interpreted as opening of fractures con-nected to the sea during the pre-earthquake deformation.
Curves C and D are records of high sensitive strain gauge installed by the Japanese Meteorological Agency at two stations in the Izu peninsula. At Irozaki(D) 40 km west-southwest of the epicenter, compressional distortion began on December 3, lasted until middle December, 1977 and then remaining flat for 20 days. After January 11, 1978, the deformation turned to tensional sense and was then followed by the main shock. At Ajiro(C) 35 km north of the epicenter, the rate of the compressional distortion had increased from December 19, which was 26 days before the main shock. There was a good conformity in the precursory behaviors of our water-level anomalies and JMA records of strainmeters(Oki and Hiraga 1978).
In the area between 50 and 150 km from the epicenter, water-

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level anomalies appeared in 26 wells beginning with 4 to 10 days before the main shock. Among them, five members reported the facts to the headquarters as marked with arrows in Figure 3. As winter is the dry season in the Pacific side of Honshu Island, the disturbance due to precipitation is not frequent and water-levels in disused wells show very smooth curves. The anomalous changes were easily identified by the deviation of water-level from the simple extension of a smooth curve. In many cases, the earthquake occurred in the final phase of the anomalies, and the water-level returned to the level smoothly extended. No pronounced coseismic disturbances in water-levels were observed in the remote area more than 50 km away from the epicenter. Yamaguchi and Odaka(1978) recognized a noticeable water-level anomaly related to this earthquake, showing conformable be-havior with JMA strainmeters as well as our water-level observation(Figure 5). The well was a disused thermal well 600 m deep at Funabara, a hot spring village in the central Izu, lo-cated 40 km west-northwest of the epicenter. The Funabara water-level corrected for barometric pressure, increased from late March to late December, 1977 with a rate of 30 cm/month. The water-level drop by 80 cm during mid-July and August was due to pumping discharge from a deep thermal well a few 100 m from the monitoring well into a swimming pool. Beginning in late December the water-level began dropping rapidly for 22 days and on January 14 the big earthquake occurred. As co-seismic behavior, the water-level dropped sharply by 7 m. The water-level behavior from late March to late December (for 9 months) is classified as the intermediate-term precursor. The water-level from late December to the main shock(for 22 days) is of the short-term precursor, comparable with the temperature rise of Kurokawa's well for 2 weeks before the quake. It is noted that the rise of water-level by 1 m in the period of the intermediate-term precur-sor returned mostly to its prior level before the quake, remind-ing us a behavior of a strain sensor.

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1980 East Off Izu Peninsula Earthquake
On June 29, 1980 a M 6.7 earthquake occurred in the sea-floor 10 km deep, 15 km east off Ito city, hot spring resort, Izu peninsula. In spite of a relatively large earthquake, the damages due to the shock were not serious. Eight persons were injured by fallen objects such as roof-tiles, rocks from cliffs along roads and collapsed window-glasses. 218 houses in the coastal area of Ito were partly collapsed. On June 25 at 18:35, the first felt earthquake swarm including several shocks with M 3 began in the sea-floor 15 km east off Ito. With intervals of several hours, many earthquake swarms repeatedly occurred and the main shock with M 6.7 occurred on June 29(Figure 6). Many after shocks were felt for a month. Twenty examples of water-level and temperature anomalies before the main quake were observed(Figure 7). On June 23 at 16:35, three days prior to the beginning of the first earthquake swarm, Y.Kurokawa telephoned to the headquarters that temperature dropped by 0.2°C from 59.0°C, which was unusually fixed for 17 days showing no reaction with barometric pressure change. As shown Figure 6, seven members telephoned Water-level anomalies to the headquarters. On June 25, the headquarters release a caution for careful monitoring to 8 selected members in the southern Kanto district. On June 25 at 18:35 several felt earthquake occurred in the Off Ito area. JMA announced that the earthquake swarms might produce several shocks, however, no large earthquake would be expected according to TV and Radio News. On June 27 at 10 to 12 o'clock, the second caution was dispatched by the headquarters. Considering the water-level anomalies which had been recognized by that time were fairly small in size, we felt that the expected earthquake would be less than M 5.5, small enough to send only a limited request for careful observation. However, it was a big shock with M 6.7. No anomaly was reported in the volume strainmeters.
1978 Off Miyagi Earthquake(M 7.4)
This earthquake, killed 27 persons, seriously damaged 2844

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houses, occurred under the sea-floor 15 km deep, 100 km east of Sendai{Figure 8). R.Suzuki(1978) in Sendai noticed that the water-level began to have an even level as compared to the general downward trend from May 24 (Figure 9). From e.arly July the response of water-level to the change of barometric pressure in her well also increased by two times. She described these facts on the formulated postcard and mailed it to the headquarters 10 hours before the quake. A similar water-level anomaly reflecting barometric pressure change more strongly appeared again. It was followed by the main M 6.0 after-shock 9 days after the main quake. The main quake also occurred in the final phase of the water-level anomaly. Behaviors almost similar to Suzuki's well occurred in other two wells in the Sendai area.
1978 Northern Tokyo Bay Earthquake
An earthquake with magnitude 4.7 occurred on August 13, 1978
in the northern part of Tokyo Bay at depth of 80 km(Figure 10). Prior to the quake, ten members around Tokyo Bay observed water-level anomalies(Figure 11) and four of them reported the fact to the headquarters, suspecting a shock big enough to feel was close (Catfish Club,1978c). We had unusually small precipitation in the Tokyo area during the 1978 summer except for the beginning of August. This small precipitation brought us very serious shortage for water supply, however, it was favorable condition for the study of the water-level monitoring due to free from effect of precipitation. Notwithstanding small magnitude 4.7 and 80 km deep earthquake, water-level anomalies associated with this quake were exceptional.
1984 Tanzawa Earthquakes
M 5.2 earthquake occurred in the Tanzawa mountains on February 14, 1984 at a depth of 25 km. No serious damage was caused. As shown in Figure 12, the epicenter was in the western margin of our monitoring network, fairly good precursors appeared in water-level and temperature of thermal water in 8 wells dis-tributed within 80 km from the epicenter (Figure 13). Y.Kurokawa

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in Usami 65 km south of the epicenter recognized again tempera-ture anomaly of a pulse type started on January 23. He reported the fact to the headquarters. A warning for careful monitoring was released to the members. As the natural yield of Kurokawa's well decreased considerably after the pulse type anomaly, pumping was made to keep the ordinary discharge of 180 1/min. This precursor occurred 22 days before the quake. The Isehara Fire Station 25 km east from the epicenter observed high water-level beginning on February 2, 12 days before the quake. Almost similar period, small water-level inflation was recognized in other 6 wells. Again, the earthquake occurred in the final phase of water-level anomalies.
1984 West Nagano Prefecture Earthquake
A magnitude 6.8 earthquake occurred in the western part of Nagano Prefecture at depth of 3 km on September 14, 1984. Im-mediately after the big shock, a huge debris avalanche from the summit of the Ontake volcano(3063 m) seriously damaged the Ohtaki village in the southern foot of the volcano and killed 29 per-sons. Large water-level anomalies appeared in two stations as shown in Figure 14. M.Mishina in Kasugai City, Aichi Prefecture, monitoring a well 20 m deep, located 80 km southwest of the epicenter recognized unusual high groundwater-level through August, 1984 in compared with her the last 8 years observation (Figure 15). Currently, her water-level decreases 30 cm/month during August. The rate of water-level decrease in August,1984 was 7 cm in spite of unusually small precipitation 27 mm/August. She telephoned the fact to the headquarters on September 3(Mishina 1984). At that moment, the headquarters were busy to release warning for careful monitoring to the members of the Kanto area, due to a temperature anomaly in Kurokawa's thermal well in Usami, followed by micro-seismic activity in the East Off Ito, Izu district. No one assumed such a destructive earthquake would occur in the foot of the Ontake volcano, 170 km away from the Izu peninsula. The Fire Station of Sakaki-Tokura-Kamiyamada

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Union, 95 km northeast of the epicenter also observed unusual in-crease of water-level from August to September, 1984, in spite of unusually small precipitation during August(Figure 16). Later, Mizutani et al.(1986) found that there were three other wells, which were monitored for ground subsidence in Komaki City, Aichi Prefecture showing large water-level increase reaching 1 m rise during the same period as Mishina recognized unusual high water-level . They applied a tank model to interpret the water-level fluctuations, and accepted the 1984 August unusual high water-levels encountered in Aichi Prefecture as precursor of the West Nagano Prefecture Earthquake. Water-level rise by 1-2 cm evaluated in the Haibara monitoring well 170 m deep, 130 km southeast of epicenter was reported as a short-term precursors, occurring 2 days before the quake(Figure 17,Geol.Surv.Japan, 1985). Mizutani et al. inferred that the water-level anomalies in the Kasugai-Komaki area 100 km from the epicenter may have been caused by the crustal deformation on the order of 10~ 7.

DISCUSSION

Magnitude, epicentral distance and duration of water-level precursor
Let us put the following two qualifications for earthquakes, which might be expected to give water-level precursors.
1. Earthquakes which struck Tokyo and Yokohama, two big cities in southern Kanto district with seismic intensity more than 3 in the JMA scale during January 1977 ,to March 1986.
2. Their depths of foci are less than 60 km.
As shown in Figure 18, 34 earthquakes are selected as the major earthquakes of central Japan in 1977 to 1986. Among them water-level precursors were observed for 16 quakes and for the rest 18 quakes no water-level precursor was observed. Figure 19 is a diagram showing a relation between the maximum epicentral dis-tance of water-level precursor and magnitude of earthquake. A line defined by the following equation gives the average distance

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limit(d km) of water-level precursors in central Japan.
log d=0.3M+0.2
This means that for M 5 earthquakes the distance limit of water-level precursors is 50 km, for M 6 is 100 km, for M 7 is 200 km, and for M 8 huge earthquakes is estimated to be 400 km. It is un-likely that earthquakes of magnitude less than 4 provide visible anomalies in shallow groundwater-level, even in the epicentral

area.

Figure 20 shows a relation between precursor time(t days) of water-level anomalies versus magnitude of the 34 selected earthquakes. The relation given by the following equation sug-gests that for M 5 earthquake water-level precursors begin 10 days before the quake, for M 6 are 21 days, for M 7 are 45 days, and M 8 are 98 days before the quake.
log t = 0.33 M - 0.65
When precursors are simply classified by length of time (30 days) as illustrated with horizontal dotted 1ine,intermediate-term precursor (month to years) by groundwater-level anomalies can be defined for earthquakes of magnitude above 6.5.
The water-level behavior of hot springs and deep drill-holes
are more sensitive than shallow groundwater. Plots of the Funabara precursor and the Ohyu geyser in Atami(1923 Kanto earthquake) together with Kurokawa's thermal well, in Usami provide longer precursor time. As indicated by Scholz et al.(1973), precursor time due to land subsidence or uplift is considerably longer than water-level behaviors.
Groundwater-level in recharge and discharge areas
To detect sharp earthquake precursors in water-level monitoring, we prefer those wells that show small, annual fluctua-tion less than 50 cm and no direct reaction from each precipita-tion. The groundwater system can be divided into the recharge area and the discharge area. The water-level amplitude for the annual fluctuation is generally large in the recharge area and small in the discharge area. In the Ashigara plain, Kanagawa

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Prefecture, for example, the annual amplitude of water-level in a 38 m well in the upstream area is as much as 13 m, and that in a 60 m well in the downstream area is 0.5 m (Figure 21). Flow pat-tern for flat, plains bounded by mountains is illustrated in Figure 22. It is seen that the water discharge in the downstream area can reflect changing of physical properties of deep grouridwater system, which might be subject to pre-earthquake strain as a whole. The excellent key wells in Usami, Ajiro and Funabara in the Izu Peninsula, and in Sendai are all located in flat plains or valley floors downstream from recharge areas. Change of barometric pressure coefficient
As described in the 1978 Off Miyagi Earthquake(M 7.4), there
was significant change in rate of water-level reaction to barometric pressure change. Barometric pressure coefficients of the water-level are classified as ordinary stage(A), precursor stage(B), and after shock stage(C)(Figure 23). Increase of the barometric pressure coeffieient(mm/mb) is considerable in the precursor stage, and very large in the after shock stage. Comparative check of water-level anomalies with record of volume strain-meters
Wideman and Major(1967) recognized that there is a definite relation between magnitude of earthquake and its associated strain release at various epicentral distances (Figure 24). Dambara(1979) proposed a revised empirical equation between an average radius(r km) of deformed area(= after-shock region) and earthquake magnitude
log r =0.51 M - 2.26
The Dambara equation illustrated in Figure 24 means that crustal deformation related with earthquakes occurs in a field of 10~ 6 to 10~ 5 strain. Superimposing of water-level anomalies on Figure 24, most of the water-level anomalies are plotted in a field of 10~ 8 ' 5 to 10~ 7 in strain release. The plots of co-seismic large water-level change such as Funabara, Usami and Sendai are dis-tributed in 10~ 7 to 10~ 6 area. The water-level anomalies plotted

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in small strain step field may suggest that the anomalies took place reversibly in elastic materials, which surround the after-shock region.
Since 1976 the Sachs-Evertson volume strain-meters have been installed by JMA in drill holes 50 to 250 m deep distributed along the coast of the Tokai and Southern Kanto district. It is curious why earthquake precursors in the Tokai and Southern Kanto district were often observed only in water-level and temperature monitorings, but not in the bore hole strainmeters except the 1978 West Off Oshima earthquake. Thanks to the summarized records of strain-meters observed at 31 stations by JMA(1986)(Figure 25),the water-level anomalies observed by the Catfish Club can be checked with them. As was referred pre-viously, the volume strainmeters at Ajiro and Irozaki both in the Izu peninsula recorded compressional strain change in the order of 5 x lO- 5 and 1 x lO' 6, respectively prior to the 1978 West Off Izu-Oshima earthquake(M 7.0). In spite of our earnest expecta-tion, however, no other sizable earthquakes occurred in the Tokai and Southern Kanto district can be correlated to the noticeable strain changes shown in Figure 25.
In the preparatory experiment conducted at each station, the volume strainmeter shows good response to the stress change by the earth tide and the barometric pressure(Suehiro 1982, Hikawa et al.1983). The barometric pressure coefficients of the strain-meters are around 10~ 8 /mb. The strain effect induced by the earth tide is in the order of 10~ 7. When we take the ratio of water column 10 cm equals 10 mb of atmospheric pressure , the increase of water-head 10 cm will be marked by the change of 10~ 7 in the strainmeter. As was described, the water-level anomalies prior to several M 5-7 earthquakes were 1 to 50 cm, which may correspond to 10- 8 to 5 x 10~ 7 in the strainmeter and almost same level with the fluctuation due to the earth tide and barometric pressure change. For this reason, the volume strainmeters installed in the Tokai and Southern Kanto district could not detect the water-

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'level anomalies below 50 cm, even if present. The JMA secular monitoring by bore hole volume strain-meters indicates that the order of crustal-strains induced by M 6 - 7 earthquakes was 10 ~ 6 to 10~ 5 in the focal area and 10~ 7 or much less in the surround-ing area. The water-level change by 1 to 10cm appeared in the surrounding area might be highly amplified signal of the pre-earthquake crustal deformation. Such small earthquake precursors might be eliminated as noises from the crude data of the strain-meters .

Co-seismic water-level change
It is emphasized that the appearance of the co-seismic water-level change is an important diagnostic feature to accept water-level anomalies before earthquakes as their precursors (Yamaguchi and Odaka 1978). We believe, however, that all the precursors are not necessarily associated with co-seismic changes. For example, the co-seismic sharp drop in the Funabara water-level as much as 7 m(Figure 5) was a catastrophic event in terms of hydrology. It could occur by the collapsing of crust in the focal and after-shock region, accumulating crustal strain more than 10~ 6. In the surrounding area for crustal-strain less than 1 0 - 7 , the crust behaves as an elastic body for stress release by earthquakes. No co-seismic change of water-level can be seen there, and the water-level moves reversibly to its prior level as shown in many examples.
Role of water-level monitoring for intermediate-term precursors Water-level precursors for destructive earthquakes based on precise monitoring are not yet adequate to give a general pattern of water-level, behavior. Excellent episodes occurred in Atami, Hot Spring Resort and Arasaki fishing port in the 1923 Kanto earthquake as good examples of the temporal behavior of intermediate-term precursors.
In Atami there was a geyser named. Ohyu, before the 1923 earthquake. The Ohyu geyser had erupted ordinarily every 5 hours (Honda and Terada 1905, Fukutomi. 1936). After December 20, 1922

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the eruption of the geyser stopped. As the geyser was a major spectacular attraction of Atami, various methods were tried to restart the eruption by a committee composed of scientists, policemen and officials of the Hot Spring Hotel Union. However, no eruption was observed until early May, 1923. On May 8 the lo-cal people was delighted to observe the eruption for 19 minutes. As shown in Figure 26, from late June the eruption had increased day by day and attractively lasted for 40 minutes, on August 31. On the next day, September 1, the destructive Kanto Earthquake occurred (Nakamura 1925). After the quake the intensity of the geyser gradually went down and finally vanished by the end of 1924. This incident beginning on December 20, 1922 about 8 months before the big quake can be treated as an intermediate-term precursor.
The other interesting episode occurred in the Arasaki fish-
ing port, located at the middle of west coast of the Miura penin-sula, 35 km east of the epicenter. The Kanagawa Prefectural Government(1927) published a comprehensive report on all dis-asters caused by the quake. Land movement related to earthquakes caused by plate motion was described as a transgressional change of seawater. There is a shrine of Kumano-Gongen on the southern point of a hill surrounding the port. Trails to the shrine had been gradually submerged under the sea for the past 100 years. When a trail was covered by seawater, a new trail would been con-structed close to the foot of the hill. The time interval be-tween trail reconstruction became shorter with years. The amount of subsidence from May of 1922 to the 1923 earthquake was reported 21 to 24 cm. At the time of the serious quake, along the eastern margin of the shalloxv bay, the old trails to the Kumario Shrine reappeared again, because of sudden uplifting of the peninsula. Figure 27 is a map illustrating the change of trails. Figure 28 shows a crustal movement inferred from the report. The rate of land subsidence during 16 months prior to the big quake was 18 cm/an, 6 times larger than the previous years.

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This rapid subsidence can also be treated as an intermediate-term p r e c u r s o r .
Mogi et al.(1986) summarized the temperature fluctuation of Kurokawa*s thermal well in Usami for the past 6 years. The ther-mal water reaches its maximum temperature as co-seismic effect just after a sizable earthquake centered nearby. This high tem-perature gradually decreases until next sizable quake, and the process may be called as an intermediate-term precursor, if necessary. As previously mentioned, several days prior to a big shock in many cases the water temperature rises in the pulse type anomalies, a short-term precursor. Mogi et al. .(1986) pointed out that a rule reduced by Kurokawa from the temperature behavior of the deep well for more than a decay was good enough to notice as precursor based on their very precise temperature monitoring. When we carefully check the episodes given in the 1923 Kanto earthquake, there is good similarity in pattern of thermal water discharge. As an intermediate-term precursor in the area sur-rounding the northern Sagami bay, natural discharge of thermal water dropped considerably possibly due to lowering of per-meability by compressional stress(intermediate-term precursor). A feivT days prior to the principal earthquake, tensional fractures began to open associated with increase of discharge and of temperature(short-term precursor).
The water-level anomalies appeared in the long distance area return to their prior levels. As the mass balance of groundwater before and after the earthquake was well kept, the behavior of these water-level changes can be referred as that of strain sen-

sors .

Many sizable earthquakes with magnitude larger than 4.5 were associated with precursors of water-level anomalies as mentioned above. Large earthquakes accompany precursors that widely dis-tributed with strong signal for long duration. Small earthquakes are preceded by precursors narrowly distributed with small signal for short duration. In small earthquakes with magnitude less than

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-1 , water-level anomalies are undetectable even iri the epieentral area. In conclusion the water-level monitoring is one of the most promising techniques for detection of intermediate- and short-term precursors of destructive earthquakes.

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Figure 1. Ground uplift in the eastern Izu peninsula between 1967 and 1976 {Geographical Survey Institute,1977).

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Figure 2 . Distribution of monitoring wells of the Catfish Club and the epicenter of the 1978 West Off Izu-Oshima earthquake. Solid circles are earthquake precursors observed (Oki and Iliraga, 1978). F:Funabara

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Figure 4. Temperature of the Usami Hot Spring(Kurokawa 1978) and groundwater-level of the Atami Fire Station(1978) showing a precursor of the 1978 West Off Izu-Oshima Earthquake(M 7.0). Two records of JMA strain-meters(1978) installed in the Izu peninsula show very good conformity in precursory behaviors with the Cat-fish Club observations(Oki and Hiraga 1978). Numbers on the water-level curve are precipitations.

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Figure 5. Change of water-level observed in a thermal well 600m deep at Funabara (F) , central Izu, showing unusual increase of the water-level until late December, followed by a 1 m drop toward the 1978 West Off Izu-Oshima Earthquake. Co-seismic sharp drop was by 7 m. (Yamaguchi and Odaka 1978) .

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Figure 6. Hourly frequency of microearthquakes (1980) occurred in the East Off Ito observed at the Hakone Volcano Observatory (Catfish Club, 1980). Temperature anomaly telephoned by Kurokawa was three days before the beginning of microearthquake activity.

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Figure 7. Summary of water-level anomalies associated with the 1980 East Off Izu Peninsula Earthquake<Cat fish Club 1980). Arrows indicate dates of anomalies reported to the headquarters.

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Distance froi Epicenter

Figure 8. Location of the epicenter of the 1978 Off Miyagi Earthquake(M 7.4) and the Catfish Club monitoring wells. Solid circles are wells showed precursors to the quake(Catfish Club, 1978b).

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1978/05 SUZUKI/REIKO/280 1978/06 SUZUKI/REIKO/280

Figure 9. Hydrograph of a shallow well in Sendai 100 km west of the epicenter showed an even water-level beginning from late May(20 days before the quake). The water-level response to change of barometric pressure increased by two times than ordinary stage from a week before the 1978 Off Miyagi Earthquake. Nine days after the main quake the responsibility of the water level to the barometric pressure increased considerably as a precursor of the main M 6 after-shock(R.Suzuki 1978). Dotted curve corrected for barometric pressure coefficient of the ordinary value{0.5 mm/mb). See Figure 23.

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Figure 10. Distribution of the water level monitoring wells of Catfish Club and the 1978 Northern Tokyo Bay earthquake(M 4.7). Solid circles are the wells of precursors observed(Catfish Club, 1978c) .

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1978 Jut.

Chiba Fire S.

Tokyo K.Haiada

Tokyo M.Maruyaia

Yokohama M.Shiiizu

Yokohama H.Tabata

Tokyo F.Takadera

Zushi H.sato
Isehara Y.Oikawa
°» S.Fukuda

Nakai K.Onoue

Tokyo F.Takadera

1978 Jul.

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Figure 11. Hydrographs showing short-term precursors to the 1978 Northern Tokyo Bay earthquake(Catfjsh Club, 1978c).

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Figure 12. The epicenter of the 1984 Tanzawa earthquake(M 5.2) and the distribution of water-level anomalies associated with the

earthquake(Catfish Club 1984a,1985).

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Figure 13. Hydrographs showing short-term precursors to the 1984 Tanzawa earthquake(Catfish Club 1984a).

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(Catfish Club)

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9 Kasugai
Hishina
(Catfish Club)
Shizuoka
0 Haibara
0 (G.S.J)
0 50 100

Figure 14. The epicenter of the 1984 West Nagano Prefecture

Earthquake(M 6.9) and the distribution of water level anomalies

associated with it(Catfish Club 1984b).

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20 301 10 20 30 1 10 20

August September
Figure 15. Hydrograph for the 1984 summer water-level observed by Mishina in Kasugai showing unusual high water-level in spite of small precipitation as compared with hydrograph of ordinary summer(Cat fish Club 198 lb,1985).

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20 301 10 20 30 1 10

August September Figure 16. Hydrograph of the 1984 summer observed by Sakaki-Tohura-Kamiyamada Fire Station, Nagano Prefecture showing unusual high water-level in spite of small precipitation as compared with 1983 and 1985 summer hydrographs(Catfish Club 1984b,1985).

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Figure 17. Water-level anomalies prior to the 1984 West Nagano Prefecture Earthquake observed at Haibara 130 km from the epicenter(Geol.Surv.Japan 1985). Marked with solid area is water-level precursor. Co-seismic drops are more pronounced.

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Jan.1977 - Mar.1986

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no precusor X

Figure 18. Distribution of major earthquakes(<60 km in depth) from January 1977 to March 1986 in central Japan. Open octagons are earthquakes for water-level precursors observed and crosses are those for no water-level precursors observed.

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(km)

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Figure 19. Average distance limit of water-level precursors ver-sus magnitude of the selected major earthquakes plotted in Figure
18. Solid circles are wells of precursors observed and crosses are those no precursors observed.

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Figure 20. Precursor time(days) of water-level versus magnitude of selected major earthquakes in central Japan in 1977 to 1986. Solid circles: water-level precursors observed

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Boundary between flaw systems of similar order
Line of force

Potential distribution on the surface of the theoretical
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Figure 22. Schematic flow pattern of groundwater system (Toth,J. 1963) .

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Figure 24. Water-level precursors plotted on the magnitude-strain step-epicentral distance diagram(Wideman and Major 1967). Water-level precursors followed by co-seismic changes occur in the strain field deformed more than 10- 7.
Solid circles: reversible water-level changes
Solid triangles: irreversible water-level changes(co-seismic changes)
Open squares: precursors observed by JMA volume strainmeters

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Figure 27. Sketch map showing change of trails to the Kumano shrine due to transgress ional change of seawater for 100 years prior to the 1923 Kanto Earthquake (Kanagawa Pref . Government, 1927) .

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Figure 28. Diagram illustrating tectonic subsidence of the Arasaki fishing port inferred from the 1927 KPG report. The rate of land subsidence for one year before the quake increased markedly by 18 cm/an , suggesting an intermediate-term precursor (Oki and Odaka 1981).

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Geochemical Precursors to Seismic Activity

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ABSTRACT

During the last three decades a significant effort has been expended to identify and quantify geophysical and geochemical changes that have been frequently reported to occur prior to seismic events in the hope of providing a means of predicting the time and location of an impending earthquake. Geochemical studies of the relationship between seismic activity and precursory phenomena have investigated whether significant changes occur in the concentrations of dissolved ions and gases in groundwaters or in the concentrations of crustal and mantle volatiles in ground gases prior to earthquakes. The results of these studies have shown that gas and ion concentration changes have occurred prior to a number of seismic events and that, in some cases, the geochemical anomalies contributed to the successful prediction of a large earthquake. However, the trends of the many of the chemical variations observed to date have not been uniform and the locations of their appearance have not been easily accounted for. There is, in addition, little agreement on the mechanisms responsible for the observed anomalies. The mechanisms that have been proposed to account for the precursory geochemical anomalies have been based on the following physical processes: ultrasonic vibration, pressure sensitive solubility, pore collapse, fracture induced increases in reactive surfaces, and aquifer breaching/fluid mixing. An analysis of the available field and laboratory data suggests that the ultrasonic vibration and pressure sensitive solubility models cannot reasonably account for a significant part of the geochemical variations observed. Although the pore collapse model could explain some of the observed chemical changes in groundwater and ground-gas, questions remain regarding its ability to generate anomalies of the magnitude observed as well as its requirement for repeated, irreversible changes in the subsurface. The increased reactive surface model appears to be able to account for the hydrogen anomalies observed as well as those associated with increased concentrations of radon. Many of the reported changes in temperature, dissolved ion and dissolved gas concentrations in groundwater are best explained by the aquifer breaching/ fluid mixing model. Analysis of the probable response of these models to the earthquake preparation process, as well as to other environmental factors, suggests that geochemical monitoring programs can provide information that may be valuable in forecasting the probability of an earthquake. However, because of the complexities and uncertainties of subsurface conditions, the absolute prediction of seismic events using geochemical methods alone does not presently appear to be feasible at our current level of understanding of earthquake preparation processes.

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INTRODUCTION

It is generally agreed that earthquakes can be preceded by several types of "anomalous" geophysical and geochemical phenomena. Beyond this, there is little agreement among the seismic research community as to the significance, mechanism, or even the definition of an "anomalous earthquake precursor." The absence of agreement regarding these precursors can be attributed in large measure to their capriciousness; each set of precursors is unique, appearing in unpredictable, and often unexplainable, locations. They typically have few common characteristics that would permit a clear determination of their underlying physical or chemical mechanisms or to predict where and when they might be found prior to the next seismic event. As a result, the number of earthquakes that have been successfully forecast using precursory geochemical (or geophysical) phenomena is most notable for its small size. None-the-less, the literature abounds with retrospective reports of chemical and geophysical changes that are attributed to subsequent seismic events; the frequency of these reports suggests that a physical mechanism does exist between the observed phenomena and the earthquake preparation process that, once understood, may permit these chemical and geophysical changes to be used in our earthquake forecasting efforts.

The present study will review reports of geochemical and hydro-geochemical phenomena that have been postulated to be earthquake precursors. The precursory anomalies will be discussed in terms of the physical or chemical processes that have been proposed to account for their relationship to earthquakes as well as with respect to other, potentially interfering, non-seismic controls over their occurrence. The chemical and physical mechanisms that have been proposed to account for the observed phenomena will then be examined in terms of their consistency with the available field and laboratory data base and an assessment will be made of their ability to account for the observed precursory variations. Finally, an assessment of the potential utility of several of the current monitoring protocols to earthquake forecasting will be offered.

The types of "hydro-geo-chemical" phenomena that have been identified as earthquake precursors spans a very broad range and includes both "macro" phenomena, such as spouting water or changing groundwater temperatures, as well as changes in the concentrations of a variety of chemical species in ground waters and ground gases. Because the mechanisms that may reasonably account for the observed changes dependent upon the medium in which they are found as well as the composition of the particular species being considered, the precursory phenomena will be considered in groups as appropriate to their chemical and physical characteristics.

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DISSOLVED ION ANOMALIES IN GROUNDWATERS

Groundwater anomalies have been among the earliest and most frequently reported phenomena to occur in conjunction with earthquake activity. Historical accounts of earthquake precursors observed in Japan, summarized by Wakita (1982), and and more recent observations in China (Jiang and Li, 1981a) reported changes in taste and temperature in wells and springs that, occurred prior to a number of major earthquakes. More sophisticated monitoring of groundwater supplies, undertaken to better characterize the extent and nature of the precursory geochemical changes, have confirmed these earlier qualitative-reports and have recorded substantial precursory changes in a variety of chemical species. The monitoring protocol most frequently employed involves discrete sampling and analysis of fluids from one or more wells at intervals of weeks to months (Barsukov et al., 1979a, 1979b, 1985; Cai et al., 1984; Chung, 1981; King et al., 1981; Li et al, 1984). Although the choice of sampling location is seldom discussed in detail, deep wells and natural thermal springs appear to be the preferred sources because they are less susceptible to influence from seasonal changes in recharge from the surface or to pumping of nearby wells. However, some monitoring programs have also included shallow water sources due to their ease of sampling and availability.

The results of these monitoring efforts have reported precursory changes in the concentrations of group I and II cations (eg. sodium, potassium, calcium, and magnesium) and anions (eg. sulfate, chloride, fluoride, and carbonate) in groundwaters at depths ranging from the surface to several thousand meters. The reported anomalies were typically associated with events having magnitudes of 4 or larger and showed both increases and decreases in ion concentrations (Barsukov et al., 1979a, 1979b, 1985; Cai, 1984; King et al., 1981). These precursory anomalies have been reported for intra-plate earthquakes in China and the Soviet Union (Barsukov et al., 1979a, 1979b, 1985; Cai, et al., 1984; Jiang and Li, 1981a; Li et al., 1985; Zhu et al., 1984), subduction zone seismic events in Japan (Wakita, 1977; 1981; 1982; 1984), and inter-plate strike slip earthquakes in the United States (King, et al., 1981). Although the trace element concentrations in ground water have been less frequently studied, elements such as mercury, radium and uranium (Barsukov et al., 1985; Chalov et al., 1977; Chung, 1981) and fluoride, lithium, strontium and barium (Cai, et al., 1984) have shown precursory concentration changes as well. It is noteworthy that several of the reported geochemical anomalies were contemporaneous with changes in groundwater flow rates and temperatures; unfortunately, however, the chemical and thermal or hydrologic changes were seldom reported for the same water
sources.

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The location and amplitude of the observed geochemical changes have been postulated by some researchers to provide an indication of the location and magnitude of the impending seismic event (Barsukov et al., 1985; Zhu et al., 1984). However, most studies have been unable to identify a useful pattern in the distribution of anomalies. In many cases water sources within the epicentral or aftershock zone showed little or no precursory change but more distant wells show substantial changes in composition (King, 1985a; King et al., 1981; Li et al., 1985). Similarly, some wells have shown significant anomalies but other near-by wells exhibited no detectable variations in chemistry or temperature. Thus there is, at present, no general agreement as to the utility of the observed geochemical anomalies for predicting how large or how close an impending seismic event will be.

The precursory changes in groundwater chemistry have been suggested by several researchers to be the result of mixing of fluids from two or more chemically disti.nct aquifers (Barsukov et al, 1985; Cai et al., 1984; King et al., 1981; Li et al, 1985; O'Neil and King, 1981; Wakita, 1984). This mechanism presumes that precursory deformation breaches, or at least reduces, hydrologic barriers between different groundwater systems thus allowing them to exchange fluids. This mechanism is substantiated by reports of changes in water levels, flow rates, isotopic compositions and temperatures that occur nearly coincidentally with the ion chemistry changes (Barsukov et al., 1985; Cai et al., 1984; Li et al., 1985; Wang et al., 1984; Wakita, 1982, 1984). Unfortunately, however, there are only a few cases where the the latter data are reported for the same water supplies in which the chemical anomalies were observed (King et al., 1981; O'Neil and King, 1981).

Other mechanisms that have been suggested to account for precursory geochemical anomalies include pressure sensitive solubility of some ions in water (Cai et al., 1984), fracture induced exposure of fresh rock surfaces to groundwater reaction (Cai et al., 1984), ultrasonic vibrations (Barsukov et al., 1985; Chalov et al., 1977; Mavlyanov, et al., 1971), and pore collapse and expulsion of pore-water into the circulating groundwater system (Barsukov et al., 1979a, 1979b, 1985; Jiang and Li, 1981b; Li et al., 1985). Each of these mechanisms is supported, to varying degrees by the available data and, in some circumstances, may make a significant contribution to a limited set of the observed chemical anomalies. A more detailed evaluation of the physical and chemical processes associated with these mechanisms, and their ability to account for the field data, will be discussed below.

GAS GEOCHEMICAL ANOMALIES

Changes in gas emissions from thermal and non-thermal wells and the discharge of odorous or flammable gases from the earth have been prominently reported in historical, as well as recent,

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accounts of pre-seismic phenomena (Gold and Soter, 1985; Wakita, 1982; Zia, 1984). Quantitative investigations of variations in the concentrations of subsurface gases have included: CO2, CH4, H2S, He, Ne, Ar, N2, H2, and Rn. Because the origin, chemical characteristics and probable association of these gases with earthquake precursory phenomena are highly variable, they will be discussed individually or in groups as appropriate.

Reactive Gases

Among the more spectacular variations in gas concentrations that have been reported are the descriptions of gas jets from dry wells, sand geysers, and large increases of gases effervescing from springs (Jiang and Li, 1981a; Li, et al., 1985). The primary components of these gases were often reported to be carbon dioxide, although methane and hydrogen sulfide were cited as probable constituents of flammable or odorous discharges respectively (Cai, et al., 1984). Less remarkable pre-seismic changes in the concentrations of carbon dioxide, methane and hydrogen sulfide have also been reported in the Soviet Union (Barsukov et al, 1985; Nersesov, 1984; Sardarov, 1981), China (Jiang and Li, 1981a; Li et al., 1985), Japan (Kawabe et al., 1981; Kawabe, 1985; Sugisaki and Sugiura, 1986), California (King et al., 1981) and Mexico (Santoyo et al., 1987). The response of the gas concentrations to impending seismic events has been highly variable. In the incidents cited in China and Mexico, precursory increases in the discharge rates, or concentrations, of C02, CH4, NH4 and H2S were reported to have occurred in groundwater or thermal wells and spring discharges prior to large seismic events. In Japan, Kawabe (1985) reported an increase in methane (normalized to argon) concentrations but, in contrast, Sugisaki and Sugiura (1986) found a decline in CH4/Ar ratios immediately prior to seismic events. Similarly, research conducted in the USSR and California have reported both increases and decreases in concentrations of dissolved C02 (Barsukov, 1985).

Several mechanisms have been invoked to account for the changes in concentrations of these gases. The decreases in gas concentration are, as noted above for the dissolved-ion concentrations, generally attributed to strain-induced breaching of aquifer boundaries and mixing of fluids having lower concentrations of dissolved gases (Barsukov et al., 1985; King et al., 1981; Li et al., 1985). Changes in water chemistry, when they are reported (King et al., 1981; Wakita, 1982, 1984), tend to substantiate this mechanism. Increases in gas flow rates or gas concentrations are also attributed to increased permeability between surface discharges and an aquifer containing gas-rich fluids. Potential reservoirs for the carbon and sulfur gases include mantle volatiles (eg. volcanic gases) (Gold and Soter, 1985; Irwin and Barnes, 1980; Santoyo et al., 1987), metamorphic production from shales or fossil carbon deposits (Cai et al., 1984; Kawabe, 1985; Sugisaki et al., 1980), or microbial activity in shallow alluvial deposits (Li et

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al., 1985). In many instances there is a close association between a particular gas and a probable source: methane and hydrogen sulfide were found in areas of known coal or petroleum deposits or organic rich shale horizons (Barsukov et al., 1985; Cai et al., 1984; Kawabe et al, 1981; Kawabe, 1985; Sugisaki et al., 1980) and carbon dioxide discharges were observed in association with limestone formations (Cai et al.,1984).

Two other mechanisms suggested to account for increased gas concentrations are the pore collapse (PC) model, suggested by King (1978), Barsukov (1979a; 1979b; 1985), and others, and micro-fracture generated gas release suggested by Wakita (1984). The PC model suggests that increasing stress prior to an impending seismic event causes the collapse of pre-existing pore volume and the expulsion of chemically distinct (gas and dissolved ion rich) pore fluids into the actively circulating hydrologic system where they are detected. The micro-fracture model suggests that trapped gases are released when pre-seismic deformation dilates grain boundaries, or fractures the mineral grains, in subsurface rocks with which the groundwaters are in contact (Sugisaki, 1978; Sugisaki and Sugiura, 1986). Although these mechanisms are unlikely to account for macro-phenomena such as surface gas discharges, either may contribute to the more subtle increases reported in carbon or sulfur gas concentrations.

Non-reactive Gases

The chemical and isotopic compositions of the inert gases in subsurface fluids have also been found to be responsive to seismic events. The gases that have been studied to date include He, Ne, Ar, Rn, and, although not an inert gas in a strict chemical sense, N2. The concentrations of these gases have been studied for precursory anomalies in both groundwater and in ground gases. In some instances, the relative concentrations and isotopic ratios of the gases have been determined in order to eliminate non-seismic interferences or to identify the origin of the gases.

Helium

Helium has been suggested as a possible tracer for the identification of hidden or buried fault systems because of its frequent presence in recently active faults (Barsukov et al., 1979a, 1982, 1985; Bulashevich and Bashorin, 1973; Bulashevich et al., 1976; Dikun, et al., 1975; Wakita, 1977, 1978, 1982). Helium concentrations have also been monitored for precursory anomalies in several countries as well, including: the Soviet Union (Barsukov et al., 1979a, 1979b, 1985; Bulashevich and Bashorin, 1973; Chalov et al., 1977); Japan (Wakita, 1982); and the United States (Reimer, 1980, 1981, 1985). These studies have reported increases in groundwater helium concentration prior to seismic events (Barsukov et al. 1979a, 1985; Chalov, 1977; Wakita, 1982) but apparent decreases in its concentration

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in ground gas (Reimer, 1981, 1985). Although some studies have suggested that the shape and duration of the groundwater helium anomalies could be used to estimate the distance and magnitude of an impending seismic event (Barsukov et al., 1979a, 1985), no other studies have been able to find such a correlation.

Investigations of the impact of non-seismic effects on helium have shown that high rates of rainfall, by saturating surface soils and inhibiting air/soil-gas exchange, increase near surface helium concentrations. Conversely, high wind speeds accelerate soil-gas exchange and decrease shallow helium
concentrations (Reimer, 1980). Groundwater helium
concentrations appear to be affected by hydrological (eg. mixing and dilution) and biological processes (eg. dilution by biogenic methane or carbon dioxide) (Sugisaki, 1978). In order to minimize such interferences, two studies normalized the concentrations of He, (as well as N2, Ne, and CH4) to Ar concentrations. The results of these investigations suggested that the He/Ar and N2/Ar ratios decrease substantially prior to some earthquakes (Sugisaki, 1978; Sugisaki and Sugiura, 1986). However, other investigators have found only relatively minor precursory variations in the rare gas ratios (Kawabe et al., 1981; Kawabe, 1985).

These studies have also found, by comparing the inert gas ratios during anomalous and normal periods, that two distinctly different reservoirs contribute to the gases present: one similar to air and the second having a crustal gas component that becomes more evident during the anomalous pre-seismic periods (Kawabe et al., 1981; Li et al., 1985; Sugisaki and Sugiura, 1986). These results were interpreted to suggest that a stored gas cap had been released by the breaching of an impermeable horizon (Li et al., 1985) or that the gases were being expelled due to pore volume collapse (Sugisaki, 1978; Sugisaki and Sugiura, 1986).

Secondary sources of variability in the inert gas abundance ratios were reported to include high rates of C02 exsolution (Kawabe et al., 1981; Sugisaki et al., 1980) as well as earth-tide effects. Sugisaki (1981) and Sugisaki and Sugiura (1986) demonstrated that a close association existed between tidal strain changes (on the order of 10 A -8) and variations in He/Ar ratios in spring discharges; these variations were attributed to varying release rates of trapped helium and argon from the rock matrix due to tide-induced strain changes.

The source of the gases being sampled was idantified in some studies using isotopic analysis: He3/He4 data from the Soviet Union (Barsukov et al., 1985) indicated that the release of helium from crustal rocks was responsible for the helium anomalies observed there. Similar analyses made by Wakita (1978) and Sano et al. (1986) found exceptionally high He3/He4 ratios indicating that the helium anomalies detected were

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derived from magmatic volatiles and thus that the earthquake activity was the result of volcanic rather than strictly tectonic processes.

Radon

Radon-222 is a radioactive daughter product of naturally occurring uranium present in rocks and soils and has a half-life of 3.825 days. Studies of the association between subsurface concentrations of radon and seismic activity have been underway for several decades. Anomalies attributed to seismic activity were first reported forty to sixty years ago (Imamura, 1947; Shiratoi, 1927; Hatuda, 1953; Okabe, 1956) although the more recent reports of a groundwater radon anomaly associated with the Tashkent earthquake have received wider attention (Ulomov and Mavashev, 1971). Since these early reports, radon monitoring programs have been undertaken in at least a dozen countries (Allegri et al., 1983; Barsukov et al, 1985; Friedman and Hernegger, 1978; Hauksson, 1981a; Hauksson and Goddard, 1981; King, 1985a, 1986; Liu et al., 1985; Santoyo et al., 1987; Tanner, 1980; Thomas, et al., 1986; Wakita et al., 1980a, 1985). Groundwater monitoring efforts have pursued a variety of protocols. The most frequently used technique has employed intermittent sampling from natural springs, shallow groundwater wells or from deeper (2000m) monitoring wells (Chung, 1985; Friedmann, 1985; Hauksson and Goddard, 1981; Li et al., 1985; Liu et al., 1985; Santoyo et al., 1987; Shapiro et al., 1985; Talwani et al., 1980; Teng, 1980, 1984; Wakita et al., 1985) have been among the most frequently employed methods. Subsequent analysis of radon in the liquid samples has been by gas-phase scintillation counting of the inert gases stripped from the individual samples or liquid/liquid extraction of radon and liquid scintillation counting (Wakita, 1982). Because of the short duration of many of the observed anomalies, however, continuous monitoring of gaseous radon has become a more accepted method in the recent past. Continuous monitoring of radon has been done using direct alpha particle counting in ZnS scintillation chambers, and beta or gamma counting of filters impregnated with the radon daughter products (Pb214, Pb210, Po214, and Bi214)(Chung, 1985; Melvin, et al., 1978; Noguchi and Wakita, 1977; Shapiro et al., 1982, 1985; Smith et al., 1976; Teng, 1984).

The results of these radon monitoring programs have shown that many earthquakes have been preceded by precursory radon anomalies. The first well-documented anomaly showed a nearly threefold increase in radon activities that peaked at the time of the 1966, M=5.3, Tashkent earthquake (Gorbushina et al.,
1973). This anomaly spanned a period of at least one year but,
because long gaps exist in the data set, may have lasted several
years. More recent studies have reported pre-seismic increases
in radon activity that have had durations of several months to more than a year and exceeded prior background concentrations by several tens of percent to a few hundred percent (Allegri et al., 1983; Friedman, 1985; Li et al., 1985; Mei, 1984; Nersesov, 1984; Santoyo et al., 1987; Wakita et al., 1985; Zhang and Fu,
1981). The majority of the reported anomalies have, however,

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been of much shorter duration, lasting for periods of a few hours to a few days and typically show increases in activity above the prior long term mean (Barsukov et al, 1979a, 1985; Cai et al., 1984; Chalov et al, 1977; Chung, 1985; Hauksson, 1981b; Hauksson and Goddard, 1981; Li et al., 1985; Liu et al., 1985; Santoyo et al., 1987; Talwani, 1981; Talwani et al., 1980; Teng, 1980, 1985; Teng et al., 1981; Wakita, 1982; Zhu et al., 1984). However, some studies have suggested that short-term decreases in radon activity (Cai et al., 1984; Talwani et al., 1981; Wakita, 1981, 1984), having durations of a few days to a few months, should also be considered to be earthquake precursors. In at least two cases (Cai et al., 1984; Santoyo et al., 1987), a decline in radon activity at one water source was contemporaneous with an increase in a second one.

The mechanisms that have been suggested to account for precursory radon variations have included ultrasonic vibrations (Cai et al., 1984; Gorbushina, et al., 1973; Mavlyanov et al., 1971), expulsion of pore fluids by collapse of pre-existing porosity (Barsukov et al., 1979a, 1985; Cai et al., 1984), release of "stored" radon due to microfracturing that accompanies dilatency (Hauksson, 198la; King, 1985a; Zhang and Fu, 1981), closing of pre-existing microfractures to inhibit the release of radon (for precursory decreases in radon)(Talwani et al., 1980), and mixing of fluids between aquifers having different concentrations of radon (Cai et al., 1984; Hauksson and Goddard, 1981; King, 1985a; Li et al., 1985; Talwani et al., 1980; Teng, 1980; Wakita, 1981). As will be discussed in more detail below, the fluid mixing and microfracturing mechanisms seem best able to account for the observed changes in radon activities in the field and are best substantiated by available laboratory data. The physical processes suggested by the other models seem to be less consistent with laboratory or field experiments or are unlikely to be able to account for the range of precursory phenomena observed.

The concentrations of radon in groundwater have also been found to be sensitive to a number of environmental effects that are not related to the earthquake preparation process. A detailed analysis of some of the more important environmental factors are presented by Shapiro et al. (1985) who used pattern recognition and Fourier analysis to demonstrate the effects of groundwater pumping, rainfall recharge rates, thermo-elastic strain changes, and barometric pressure changes. Fourier analysis indicated that two other effects, having periods of O.995 days and 3.82 days, were also modulating the radon signal; although no clear mechanism was proposed for these signals, the former is very close to the tidal period (0.997d) and the latter to the half-life of radon-222 (3.825d). Several other studies have substantiated the results of Shapiro et al.: groundwater pumping was identified in several localities as either increasing or decreasing radon activities (King, 1985a; Teng et al, 1981; Wakita et al., 1985). Increased rates of rainfall were found to depress the radon activities under some circumstances (Cai et

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al., 1984; Taiwan! et al., 1980; Teng et al., 1981; Wakita et al., 1985) but, if the monitoring well was of sufficient depth or was hydraulically isolated from surface aquifers, rainfall did not exert a significant influence over radon activities at depth (Wakita et al., 1985). Barometric pressure changes were found to have a relatively modest effect on groundwater radon in most cases (Wakita, 1984; Wakita et al., 1985). However, one study found that effervescence of carbon dioxide from well waters at reduced barometric pressures stripped radon from the well waters resulting in large and erratic variations in the subsurface radon activities (Shapiro et al., 1982, 1985). Several studies have identified a strong correlation between diurnal radon variations and tidal strain changes (Wakita, 1984; Wakita et al., 1985), however other investigators have found their respective frequencies to be sufficiently different to exclude a tidal influence (Friedmann, 1985; Shapiro et al., 1985).

Although recent research has emphasized the analysis of radon variations in groundwater, an association between seismic activity and the concentration of radon in soil gas was identified more than forty years ago when Hatuda (1953) found that a substantial increase in ground-gas radon activity was generated by the great Tonankai earthquake that occurred in April, 1944. More recent studies of ground gas radon concentrations and seismic activity have been conducted in the Soviet Union (Abduvaliyev et al., 1984), Japan (Yamauchi and Shimo, 1982), the United States (Birchard and Libby, 1980; Cox, 1983; Cox et al., 1980; Fleischer and Mogro-Campero, 1985; King, 1978, 1980a, 1980b, 1985b; Mogro-Campero et al., 1980; Steele, 1981, 1985; Talwani et al., 1980; Thomas et al., 1986; Thomas and Koyanagi, 1986), and, for fumarolic discharges, Iceland (Hauksson, 1981b). The majority of these studies have used nitrocellulose films (Cox et al., 1980; Fleischer and Mogro-Campero, 1978; Steele, 1981, 1985; Talwani et al., 1980; Thomas et al., 1986) or, more recently, less temperature sensitive polymer (allyl diglycol carbonate) chips (Fleischer and Mogro-Campero, 1985) to measure an integrated average radon concentration present in the soil gases at depths of 30cm to 100cm. This method is inexpensive and quite versatile in terms of the monitoring location but the temporal resolution possible varies from one to three weeks depending upon the soil radon
concentrations. Sampling frequencies on the order of hours have
been achieved in other studies using ionization chamber pulse counting devices (Yamauchi and Shimo, 1982) or field portable electronic semiconductor detection equipment (King, 1985b; Thomas and Koyanagi, 1986). Gamma counting of radon daughter products, using sodium iodide scintillation counters, has also been employed to determine long-term average distributions of subsurface radon (Fernandez, 1983).)

The most frequently found precursory ground gas radon anomalies have been increases in activities prior to, or contemporaneous with, seismic events. The duration of the radon anomalies were

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highly variable depending upon the technique used to monitor the activities: most integration periods ranged from one to three weeks and hence the minimum duration was constrained to this period. In studies of hourly average activity variations, several short "spike-like" anomalies were detected that were correlated with seismic or creep events (King, 1985b). The maximum duration of anomalous periods was as long as several months (Fleischer and Mogro-Campero, 1985; Thomas, et al., 1986) although, in the latter case, this duration was not well constrained.

The amplitudes of the soil gas radon anomalies have ranged from a factor of two increase above prior backgrounds (Abduvaliyev et al., 1984; King, 1978; Steele, 1981, 1985; Talwani et al., 1980) to increases of factors of ten or greater (King, 1978, 1985b; Mogro-Campero et al., 1980; Thomas et al., 1986). In one case (Fleischer and Mogro-Campero, 1985) , decreases in radon activity by factors of more than 500 were attributed to seismic effects. The observed changes in the radon concentrations were, to some extent, dependent upon the integration period: hourly integration periods showed "spike-like" anomalies that exceeded the prior averages by factors of five to ten but longer integration periods typically showed broader, less intense, increases.

The anomaly amplitudes have also shown a general decline with increasing distance between the respective monitoring station and the epicenter of the earthquake and with decreasing magnitude of the event (Birchard and Libby, 1980; Fleischer and Mogro-Campero, 1985; King, 1978; Thomas et al., 1986). This correlation has been far from universal, however: King's (1985b) study detected short duration radon pulses that exceeded background by a factor of five for an event that occurred at a distance of 40km but, for an earthquake of nearly the same magnitude at a distance of only 20km, a spike of only twice the background levels was found and four events, occurring at distances of 3km to 12km, had no detectable accompanying anomalies. Several other studies have encountered similarly non-linear precursory radon responses (Fleischer and Mogro-Campero, 1985; Steele, 1985).

The period between the precursory anomaly and the associated earthquake generally ranged from several days prior to the earthquake in question (Abduvaliyev et al., 1984; Fleischer and Mogro-Campero, 1985; King, 1978, 1985b; Santoyo et al., 1987; Talwani et al., 1980; Thomas et al., 1986) to as long as several months (Birchard and Libby, 1980; Steele, 1981, 1985). In some cases anomalies occurred contemporaneously with, or after, the events (Birchard and Libby, 1980; King, 1985b; Thomas et al., 1986).

Volcanic phenomena have also been found to have a significant influence over subsurface radon concentrations. King (1978) reported a correlation between soil gas radon concentrations and

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seismic activity associated with an eruption at Kilauea volcano. More recently, other investigators (Cox, 1983; Cox et al., 1980; Thomas et al., 1986; Thomas and Koyanagi, 1986) reported significant increases in radon activity in shallow ground gases associated with several eruptions on Kilauea. Del Pezzo et al. (1981), in a study conducted on Vulcano Island (Italy), found volcanically induced radon activity increases in shallow, non-thermal, groundwater supplies and Chirkov (1975) and Hauksson (1981b) reported increases in fumarolic radon concentrations at Karymsky (USSR), and Krafla (Iceland) respectively. The Vulcano study showed precursory increases in radon concentrations but the latter two studies reported variations that were contemporaneous with the volcanic events.

The most frequently proposed mechanism to account for the pre-seismic increases in soil-gas radon activity has been that regional changes in the strain field accompany, or precede, seismic events and that these changes either compress or dilate the subsurface pore volume. Compression is believed to force radon rich ground gas from depth toward the surface soil layer, resulting in an increase in radon activity, whereas dilation draws radon-poor atmospheric air into the soil and thus decreases its radon concentrations (Fleischer and Mogro-Campero, 1985; King, 1978, 1980a, 1985b; Mogro-Campero et al., 1980). Other mechanisms that have been proposed to account for the earthquake induced radon increases have been the release of deeper crustal gases by a precursory dilation of fracture systems (Abduvaliyev et al, 1984) and the local generation of radon by strain induced microfracturing and release of intergranular radon (Thomas et al., 1986). Mechanisms proposed to account for volcanically induced radon increases have included changes in water level (Hauksson, 1981b), increased subsurface steam discharge (Cox et al., 1980; Gasparini and Mantovani, 1978), and strain induced release of radon from the subsurface rocks (Thomas et al., 1986).

Comprehensive reviews of natural variations of radon activities in soil gases, in addition to those associated with earthquake or volcanic activity, have recently been published by Tanner (1964, 1980). One of the more important results of the work reviewed is that mixing between radon deficient air and radon rich soil gas has resulted in a steep concentration gradient of radon in most shallow ground gases (Fernandez et al., 1983; Finck and Persson, 1981; Kraner et al., 1964). Soil permeability substantially affects the exchange rate of air with subsurface gases and thus the soil morphology as well as such environmental factors as frozen or rain saturated ground have a significant influence over the spatial and temporal variations in soil radon concentrations (Fleischer, 1983; Guedalia et al., 1970; Klusman, 1981; Klusman and Webster, 1981; Kovach, 1946; Kraner, et al., 1964; Megumi and Mamuro, 1973; Yamauchi and Shimo, 1982). Soil moisture also influences radon emanation from individual soil grains due to recoil-implantation and diffusional effects (Fleischer, 1983; Tanner, 1980). Diurnal

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and synoptic barometric pressure changes have the effect of "pumping" ground gas: increasing pressure forces radon deficient atmospheric gases into the ground, lowering radon activities, and decreasing pressure permits radon rich ground gas to "exhale" from depth thus increasing radon concentrations in the shallow soil layer (Hatuda, 1953; Klusman, 1981; Klusman and Webster, 1981; Kraner et al., 1964; Schery et al., 1982). Other environmental factors that influence radon activities in ground gas include wind speed and turbulence (Guadelia et al., 1970; Israelsson, 1980; Kraner et al., 1964; Pearson and Jones, 1965) and ground/air temperatures and temperature differences (Klusman, 1981; Klusman and Webster, 1981). There is, however, no general agreement as to the relative importance of these effects in a radon precursor monitoring effort; whereas some studies have clearly shown diurnal and meteorological effects (Fleischer and Mogro-Campero, 1978; King, 1985b), others have found none (Fleischer and Mogro-Campero, 1985; Thomas et al., 1986). Although these differences may be the result of variations in monitoring protocol or in local soil and meteorological conditions, their occurrence suggests that meteorological effects need to be considered in a precursory monitoring effort.

Hydrogen

The association between hydrogen gas concentrations and fault movement has been recognized only relatively recently. Sugisaki et al. (1980), in a comprehensive analysis of fault gases in the vicinity of Nagoya, Japan, found highly variable concentrations of hydrogen, and other gases, within and between faults depending on rock type and on fault activity. Wakita et al. (198Ob) have reported extremely high enrichments of hydrogen at the Yamasaki Fault relative to concentrations typically found in either atmospheric or soil gases. The coincidence of high concentrations of hydrogen with an active fault zone led both groups to postulate that there was a genetic relationship between fault activity and hydrogen generation. More recently, Sugisaki et al. (1983) reported a general correlation between the age of most recent fault activity and the degree of hydrogen enrichment. However, they also found that the concentration of hydrogen could vary by orders of magnitude from place to place on, or adjacent to, a fault and suggested that the rock type and degree of alteration could substantially affect the concentrations of hydrogen present regardless of the age of activity. This finding was confirmed by Ware et al. (1985) who conducted hydrogen surveys on several fault systems, both inactive and very recently active, and found little correlation between activity and hydrogen content. The latter study indicated that both hydrothermal activity and oxidation of iron hydroxide were capable of generating strong enrichments of hydrogen under field conditions. In a related study, isotopic analysis of hydrogen in fault gases (Kita et al., 1980) indicated that its depth of origin was 20km, or possibly much less, thus suggesting a crustal origin for the hydrogen.

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The application of hydrogen as an indicator of impending seismic activity has been investigated by researchers in Japan (Nagata et al., 1979; Satake et al., 1985; Sugisaki and Sugiura, 1986; Sugisaki, 1985), China (Li et al., 1985), and the United States (O'Neil and King, 1981; Sato and McGee, 1981; Sato et al., 1985; Sato et al., 1986). Many of the monitoring programs have employed discrete sampling of gas from groundwater or ground gas and subsequent analysis by gas chromatography; however, more recent studies have monitored hydrogen concentrations on a continuous basis using an adaptation of hydrogen fuel cell technology (Sato and McGee, 1981; Sato et al., 1986).

Many of these studies have reported increases in hydrogen concentrations either prior to, or contemporaneous with, seismic activity. The discrete sampling methods have shown variable hydrogen enrichments and, as with all other geochemical anomalies, there is little agreement regarding the timing or amplitude of the hydrogen anomalies and the seismic events that appear to have generated them: Sugisaki (1985) finds a general correlation between the hydrogen concentrations observed and earthquake magnitudes along a particular fault but Satake et al. (1985) find a more variable hydrogen response to both the magnitude and intervening distance of seismic activity. Similarly, Sato et al. (1985, 1986) have reported variable hydrogen responses to both seismic and creep events along the San Andreas Fault: some creep and earthquake events show strong hydrogen anomalies whereas others show none. Some of the latter variability in the response may, however, be the result of instrumental difficulties associated with advancing the state of the art for hydrogen monitoring instrumentation.

Three processes have been proposed to account for precursory releases of hydrogen prior to earthquake activity. Li et al. (1985) suggest that hydrogen is expelled from pore spaces due to increasing strain prior to a seismic event. Sato et al. (1985, 1986) have attributed the increase in hydrogen activity along the San Andreas Fault to the serpentinization of ophiolitic rocks at depth beneath the fault. Because there is an increase in volume associated with this process, the fault system is dilated thus generating an earthquake and releasing hydrogen from depth. The third process suggests that a chemical reaction between water and silicate free-radicals on fresh rock surfaces, exposed by precursory microfracturing, generates hydrogen gas (Kita, et al., 1980; Sugisaki, 1985; Wakita et al., 1980b). The latter process has been demonstrated in laboratory experiments that showed that crushed rocks, when placed in contact with water, could generate substantial quantities of hydrogen. This study also found that unaltered rocks produced more hydrogen than highly altered fault gouge material (Sugisaki et al., 1983) . Recent spectroscopic analyses of light generated during rock fracturing has helped substantiate this mechanism by showing that molecular hydrogen is generated when granites or basalts are fractured in a water saturated environment (Brady and Rowell, 1986) . Additional support for the "local"

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generation of hydrogen by rock-water reactions comes from the isotopic evidence cited above (Kita et al., 1980) as well as that of O'Neil and King (1981) who showed that there was a significant deviation of the D/H isotopic composition of the water phase that was contemporaneous with other precursory groundwater chemical anomalies. The D/H change in the latter study suggested that hydrogen had been lost as a result of a water-rock reaction.

ANALYSIS OF MECHANISMS RESPONSIBLE FOR GEOCHEMICAL ANOMALIES

The results of the monitoring programs described above clearly suggest that several types of geochemical anomalies can precede seismic activity. More importantly, some researchers have reported very impressive successes in predicting what might otherwise have been tragic earthquakes. However, there have also been many notable failures to predict devastating seismic events as well. Furthermore, close inspection of some of the data upon which the successful predictions were allegedly based, has found inconsistencies and uncertainties that have led to skepticism of the utility of geochemical monitoring in general. Even though it is evident that the record is still far from complete, there may now be sufficient data available with which to form a general, although tentative, model of the mechanisms responsible for some of the precursory geochemical phenomena observed.

Although many nominally different mechanisms have been suggested to account for precursory geochemical anomalies, most are based on the following physical or chemical processes:
1.) Physio-chemical release by ultrasonic vibration
(UV model);
2.) Chemical release due to pressure sensitive solubility
(PSS model);
3.) Physical release by pore collapse (PC model);
4.) Chemical release by increased loss from, or reaction
with, freshly created rock surfaces (IRS model)
5.) Physical mixing due to aquifer breaching/fluid mixing
(AB/FM model).

Each of these processes is supported, to some degree, by the available field or laboratory data. However, a careful examination of each, in terms of its ability to account for the reported anomalies and its consistency with what is known of the earthquake preparation process, suggests that some are much better able to account for the observed geochemical precursors than are others.

Ultrasonic Vibration Model

The ultrasonic vibration model has been invoked to explain increased concentrations of a variety of dissolved ions and gases in groundwater and changes in the relative concentrations of ground gases (Barsukov et al., 1979a, 1979b, 1985; Chalov et

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al., 1977; Mavlyanov et al., 1971). This model is based on the theory that a variety of loosely bound elements or compounds present in subsurface rocks can be mechanically freed by ultrasonic vibrations occurring prior to seismic events. The theory is substantiated to a degree by laboratory studies that have shown that rock samples exposed to ultrasonic treatment have lost some ions and gases more rapidly than samples not exposed to such treatment (Barsukov et al., 1985). Other investigations (Barsukov et al., 1985; Cai et al., 1984) also report field studies in which explosive discharges resulted in increases in the concentrations of various compounds in groundwater at some distance from the point of discharge.

On closer examination, however, this mechanism appears to be able to contribute little to the changes observed. Although there is little doubt that ultrasonic treatment of rocks under a laboratory setting can increase rates of reaction and degassing, it can be argued that this increased reactivity is the result of cavitation and mixing of fluids at the rock-water interface at extremely high rates of energy input. Comparison of these levels of energy with field data on the spectra of micro-earthquakes that could conceivably generate this ultrasonic vibration indicates that the frequency spectrum of these acoustic emissions are predominantly in the range of 1.5kHz to 6kHz (Teng, 1984; Teng and Henyey, 1981). The high frequency component of these micro-earthquakes is far too weak to contribute significantly to the release of ions or gases from subsurface rocks. Similarly, explosion-induced changes in groundwater chemistry were, at best, modest in comparison with the naturally occurring precursory changes. Furthermore, the several hour delay reported between the explosions and the observed increase in ion concentrations (Barsukov et al., 1985; Li et al., 1984), suggests that some phenomenon other than simple vibrational release was responsible for the observed increases in ions and gases.

Thus, even though laboratory data show that rocks have released additional gases and ions when exposed to ultrasonic vibration, it is clear that the energy imposed on these rock samples in an artificial setting is far higher than that which could naturally occur in the field prior to seismic activity without itself being easily detected. Furthermore, in those instances where near-surface fluids were exposed to strong vibrational impulses under normal field conditions, the observed chemical changes occurred long after the vibrations ceased. Thus, it is considered unlikely that this mechanism can reasonably account for the geochemical precursors that have been observed.

Pressure Sensitive Solubility

It is well established that the solubility of some ionic species and nearly all gases are sensitive to the pressure to which the fluids are exposed (Helgeson, 1969; Kharaka and Barnes, 1973). It has been suggested that precursory changes in the stress

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field associated with the earthquake preparation process are sufficient to cause an increase in the solubility of a variety of dissolved ions in groundwater and thus produce the geochemical changes observed (Cai et al., 1984). An analysis of the hydrologic anomalies that accompany stress changes suggests, however, that this mechanism is unlikely to contribute significantly to changes in fluid chemistry. The pressures needed to change the solubilities of most ions in an aqueous phase by a significant factor (eg. 50%) are on the order of tens to hundreds of bars. Even though a stress change on this order may be associated with rock fracture, similar pressure changes would have to be transmitted to the water phase before the solubilities of the pressure sensitive species could react to them. Changes of these magnitudes are not evident in the literature; the largest variations in groundwater heads and flow rates reported have only been on the order of a few bars at most (Cai et al., 1984; Wang et al., 1984) and many water sources reporting chemical changes have reported much smaller head changes. Even though the effect of small pressure changes on gas concentrations in groundwater would be more pronounced, many of the dissolved gas anomalies reported involve gases (H2, He, Rn) that are well below saturation even during the anomalous periods. Hence, the changes in the hydrostatic pressure that do occur are not likely to significantly affect their concentrations. On this basis, then, it is believed that the effects of pressure sensitive solubility on dissolved ion and gas concentrations will be very small.

Pore Collapse Model

The pore collapse model has been suggested to account for several types of anomalies including increases in the concentration of dissolved ions in groundwater (Barsukov et al., 1979b; 1985; Jiang and Li, 1981b; Li et al., 1985), changes in the absolute and relative concentrations of dissolved gases in groundwater (Barsukov, et al., 1979b, 1985; Sugisaki, 1978; Sugisaki and Sugiura, 1986), and increases in radon concentrations in shallow ground gases (Fleischer and Mogro-Campero, 1985; King, 1978, 1980b, 1985b; Mogro-Campero et al., 1980). The basic rational for this mechanism is that, at increasing stress levels prior to a seismic event, pre-existing pore volume begins to collapse, thus expelling pore fluids. It is suggested that these chemically distinct pore fluids, by mixing with the circulating groundwater system, generate the geochemical anomalies observed. Similarly, the collapse of pore volume in the unsaturated zone, by generating an outflow of ground-gas, is believed to cause and upward displacement of the normal radon concentration gradient in the shallow soil layer.

Laboratory evidence supporting the loss of pore volume has been found in a number of studies (Brace, 1977, 1978a, 1978b; Byerlee, 1978; Zoback and Byerlee, 1974) that have shown that pore volumes can decrease in some geologic materials when placed under stress. Those materials that show the greatest changes

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are unconsolidated materials and fault gouge. Tests done on sandstone (Brace and Riley, 1972) and tuffs (Honda et al., 1982) have also shown that significant losses of pore volume can occur in more competent rocks as well.

Field evidence that seems to substantiate this mechanism was reported by Sugisaki (1978) and Sugisaki and Sugiura (1986) who identified small cyclic changes in inert gas ratios that were correlated with tidal strain changes. Other studies, of explosion induced radon variations in groundwater, reported a decline in the amplitude of the anomalies with increasing numbers of repeated events (Barsukov, 1985). The latter data suggests that the locally produced radon has been gradually exhausted from the pre-existing pore volume; if radon were being generated by mixing or by release due to freshly generated fracturing (see below), it is likely that the supply would not be easily exhausted.

The significance of the PC model to precursory earthquake phenomena is, however, open to question on several counts. The laboratory data reviewed by Brace (1978a), and others, indicate that the loss of pore volume in porous rocks is essentially irreversible. Furthermore, losses in most competent rocks are quite small and occur at relatively low stress levels. At stress levels approaching the failure strength of the rock, pore volume increases in most rocks and can, in some cases, exceed that initially present. In addition, other studies of gas emanation from rocks under stress show that, at the low to intermediate stress levels associated with pore closure, gas emanation declines due to closure of the channels through which the gases escape from the rocks (Holub, 1981; Holub and Brady, 1981). It is only at the higher stresses that gas emanation increases appreciably (see also Giardini et al., 1976 and Honda et al., 1982). Because these increases are accompanied by acoustic emissions associated with micro-fracturing (Holub, 1981; Holub and Brady, 1981; Sobolev, 1984; Varshal et al., 1985) it is believed that the gas losses are the result of new pore volume formation associated with dilatency (see below). Based on these findings, the pore collapse model proposed to account for increases in ion and gas concentrations in groundwaters is considered to be a possible cause of increases in concentrations of dissolved species although it is not believed to be a major factor in the precursory anomalies observed.

The application of the PC model to the precursory radon variations observed in ground gas, although subject to the same criticisms noted for groundwater anomalies, are questioned on additional grounds as well. The short half-life of the Rn-222 restricts its radius of travel in the soil column under normal conditions and hence the concentration of radon found in shallow surface soils reaches an equilibrium concentration within a few meters of the soil surface. As noted above, near-surface soil gas radon is lost by diffusion to the air column and is diluted

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by mixing with air. The PC model suggests that the expulsion of gases from subsurface rocks generates an upward flow of gas that displaces the equilibrium soil radon profile and, as a result, radon monitoring instruments in the shallow soil detect an increase in radon activities. Closer examination of soil radon data for normal and anomalous conditions indicates, however, that this mechanism cannot be responsible for many of the documented precursory radon anomalies. Studies by Kraner et al. (1964) and more recent investigators (Clements and Wilkening, 1974; Fleischer and Mogro-Campero, 1978; Mogro-Campero et al., 1980; Tanner et al., 1980) have shown that radon profiles in most soils approach equilibrium exponentially and reach maximum concentrations at depths of only two to three meters. At depths of 0.5m to 1.0m, under no-flow conditions, radon activities are typically nearly half the equilibrium concentrations. Thus, upward flow of ground gas, regardless of rate, can only increase the detected radon concentration by a factor of two or three. However, several investigators (Fleischer and Mogro-Campero, 1985; King, 1985b; Thomas et al, 1986; Thomas and Koyanagi, 1986) have detected precursory anomalies, at depths of 0.5m to 1.0m, that were higher than long-term averages radon activities by factors of five to ten. It is, therefore, considered unlikely that pore collapse, and resultant upward migration of ground-gas, can, by itself, account for such large radon increases.

Increased Reactive Surface Model

Several investigators have proposed that micro-fracturing prior to major seismic events could be responsible for the increases in ion and gas concentrations observed in subsurface fluids (Cai et al., 1984; Hauksson, 1981a; Sugisaki et al., 1983; Thomas et al., 1986; Wakita, 1982). This model suggests that precursory stress changes generate extensive micro-fracturing in the subsurface. The fresh silicate surfaces exposed by this fracturing permit both an increased rate of reaction with the ambient groundwater and also allows the escape of trapped gases to either the groundwater or ground gas system.

Laboratory studies of whole or jointed rocks have clearly established that, at stresses approaching failure strengths, dilatency and pervasive micro-fracturing occur. Pre-failure increases in porosities have ranged from 20% to as high as 400% in competent rocks (Brace, 1977, 1978a, 1978b; Byerlee, 1978; Sobolev, 1984; Sobolev et al., 1984; Wang et al., 1978; Zoback and Byerlee, 1974, 1975). Dissolution or alteration of the fresh rock surfaces created could significantly increase ion concentrations present in groundwater exposed to the newly generated surface area; such increases would be especially noticeable in well-established systems where older rock surfaces would be effectively passivated by equilibrium alteration assemblages. Studies of the response of occluded gases in whole rocks to changing stress conditions, also support this mechanism: gas emanation rates are low at low stress levels,

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but, at progressively higher stresses, when acoustic emission associated with micro-fracturing begins to occur, the quantities of gas released increase substantially.

Field data that substantiates the association between increases in the concentration of radon (and other gases) and changes in the regional stress levels is that of Wakita et al., (1985) who showed that there was a very close correlation between the trends in regional strain, as measured by volumetric strain meters, and trends in groundwater radon concentrations. Similar data have been found by Thomas and Koyanagi (1986) who detected cyclic changes in radon activities that were closely correlated with stress changes associated with the inflation-deflation cycle at Kilauea volcano.

Although variations in the concentrations of such gases as radon or argon may be explained by changes in the rate of physical release of these gases, hydrogen anomalies are considered to be the result of chemical reactions occurring on fresh silicate surfaces. Several laboratory studies have shown that freshly created rock surfaces are capable of producing significant quantities of hydrogen through the reaction of water molecules with silicate radicals created by the fracture process (Kita et al., 1980; Kita and Matsuo, 1982; Sugisake et al., 1983). Observations of hydrogen spectra generated by fracturing rocks further substantiate this mechanism (Brady and Rowell, 1986).

Other mechanisms that have been proposed to account for the observed hydrogen anomalies do not seem to be as well supported by the available evidence. Oxidation of iron present in subsurface rocks, suggested by Ware et al. (1985), may explain long lived sources of fault hydrogen but, unless fracture-mediated exposure of fresh iron bearing minerals is invoked, this mechanism cannot easily account for the episodic nature of some of the hydrogen anomalies observed (Sugisaki and Sugiura, 1986). The model proposed by Sato (Sato et al. 1985, 1986), of hydrogen release due to dilation of the San Andreas fault by serpentinization of deep crustal rocks, is also difficult to accept on several grounds. Because hydrogen has a very high diffusion rate through most materials, its migration from depths of ten kilometers, or more, to the surface should result in very wide-spread and long-lasting anomalies. However, the hydrogen anomalies that have been detected on this fault generally show relatively short term increases that are often preceded and followed by undetectable concentrations. Thus the field data do not seem to support the hypothesis of hydrogen release from depth.

However, a significant question still remains unanswered with regard to this increased reactive surface model. This is whether the precursory stress changes, and their resultant strain, would be able to generate enough micro-fracturing in the far field to produce the observed anomalies. As noted above, several laboratory studies have shown that micro-fracturing and

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pore volume increase typifies the response of many types of rocks to increasing stress. Furthermore, rock dilatancy is a key element in at least one model of the earthquake preparation process (Mogi, 1977; Rice and Rudnicki, 1979; Scholz, 1987; Scholz et al., 1973). However, the dilatancy theory, as presently accepted, applies only to a volume of rock that is in the immediate vicinity of the fault that is broken by the earthquake (ie. the asperity volume). There is at present no suggestion that the dilatant zone could extend out from the asperity volume to distances of tens to hundreds of kilometers where many of the geochemical anomalies have been reported. Hence, some mechanism other than direct stress failure is needed to account for increased micro-fracturing and the creation of new rock surfaces that the IRS model requires.

A mechanism that might be able to account for the far-field effects without requiring high stress levels has been suggested by Anderson and Grew (1977). This model suggests that stress corrosion cracking could contribute significantly to crack propagation at stress levels well below normally accepted failure strengths. Hence, relatively small changes in stress levels could significantly accelerate or retard stress corrosion mediated micro-fracturing. Although further studies, that might attempt to detect a correlation between acoustic emissions associated with micro-fracturing and increased rates of radon or hydrogen release should be done to better confirm Increased Reactive Surface model, the available evidence seems to suggest that it could contribute significantly to the precursory dissolved-gas and ground-gas anomalies observed. Its contribution to precursory changes in dissolved ion chemistry is, however, unlikely to be significant because of the slow reaction rate of most minerals with groundwaters.

Aquifer Breaching/ Fluid Mixing Model

As noted above, the Increased Reactive Surface model is unlikely to be able to account for increases in groundwater ion concentration of more than a few milligrams per liter during the relatively limited duration of the anomalies. More importantly, it clearly cannot account for the decreases in ion concentrations that have been observed prior to a number of seismic events (Barsukov et al., 1985; King et al., 1981; Wakita, 1982). A mechanism that can account for a decline in ion concentrations suggests that mixing of groundwaters between different aquifers dilute the fluids being monitored and thus generate a precursory decrease in the ion concentrations found in the the sampled aquifer. The intruding fluids could also generate substantial increases in ion and gas concentrations as well as such other anomalies as temperature changes or variations in discharge rates or long-term changes in water levels.

In order to demonstrate mixing of fluids from distinctly different aquifers, it is necessary to have a broad suite of

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chemical analysis for such ions as sodium, potassium, calcium, and silica or concurrent analyses of temperature, chemical, and isotopic data. This type of analysis would allow one to demonstrate mixing by showing changes in equilibrium temperatures of the fluids using, for example, Na-K-Ca or silica geothermometers, or by showing differences in the source region of the water by changes in the deuterium/hydrogen and oxygen-18/oxygen-16 ratios. Unfortunately, this type of investigation is the exception rather than the rule. One such study, however, done by King et al. (1981) and O'Neil and King (1981), showed that concurrent changes in chemical and isotopic compositions of fluids from several wells were clearly the
result of groundwater mixing. Several other monitoring
programs, although not reporting full data sets, have shown that changes in chemical composition, temperature, and level or flow rate of groundwater often occur concurrently (Barsukov et al., 1985; Jiang and Li, 198la, 198Ib; Li et al., 1985; Mei, 1984; Wakita 1981, 1982, 1984; Zhu et al., 1984). The reported anomalies include examples of falling chloride and cation concentrations that are accompanied by decreasing temperatures and increasing flow rates or water levels (Cai et al., 1984; Jiang et al., 1981a; Li et al., 1985), or declining radon concentrations that are accompanied by falling temperatures (Wakita 1982, 1984; Wakita et al., 1980a). Hence, the chemical and hydrologic observations strongly suggest that mixing of groundwaters of distinctly different characteristics is the predominant effect generating many of the decreases in ion concentrations as well as the substantial increases in dissolved ion concentrations observed.

The physical mechanism responsible for the mixing of fluids is not as clearly understood. Laboratory studies of whole-rock samples under varying levels of stress have, as indicated above, shown that the porosities of a number of rock types increase by as much as several hundred percent when loaded to 80% to 90% of their failure stress (Brace, 1977, 1978a, 1978b; Byerlee, 1978; and Zoback and Byerlee, 1975). However, differences between resistivity and permeability changes in the stressed samples have been interpreted to indicate that most of the fracture porosity created is not interconnected (Zoback and Byerlee,
1974, 1975). More-over, the absolute increase in matrix
permeability is still very small (Brace, 1978a, 1978b). Hence the observed stress induced increases in bulk porosity and permeability may have only a minor influence over groundwater
flow. More significant effects on permeability have been found
in investigations conducted on jointed or artificially saw-cut samples. These studies have reported that fluid flow through existing fractures could be increased or decreased by several hundred percent by relatively small changes in stresses normal to, or parallel to, the joint face (Brace, 1978a, 1978b; Byerlee, 1978). The permeability response of clays or other fault-gauge materials to changes in stress, and their impact on these results is, as yet unclear (Brace, 1978a; Byerlee, 1978).

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Even though the above mechanism is very attractive in terms of its consistency with laboratory and field data, the stress changes that occur at distances of tens to hundreds of kilometers away from an impending epicenter, where the geochemical anomalies have appeared, are believed to be quite small. Furthermore, the question of how large scale mixing of fluids between aquifers could be effected without massive groundwater displacement is a significant obstacle to the acceptance of this mechanism. Thus significant questions still remain regarding the ability of far-field stress changes to effect broad scale fluid mixing between separate aquifers.

An alternative physical mechanism that could be considered to account for changes in the fluid compositions discharged form wells and springs does not require large-scale mixing of fluids. This model is based on changes in hydraulic head that have been observed in wells prior to, or concurrent with, some earthquakes (Roeloffs, 1987; Roeloffs and Bredehoeft, 1985; Roeloffs and Rudnicki, 1985, 1986). If a well or spring derives fluids from two or more separate aquifers having different permeabilities, temperatures, and fluid compositions, then the physical and chemical composition of the mixed fluids discharged will be critically dependent upon the hydraulic head within each aquifer. A change in the relative hydraulic heads among the aquifers, brought about by a precursory stress change, could therefore have a substantial impact on water chemistry without requiring large scale mixing of fluids away from the fluid discharge conduit. An attractive feature of this model is that it could also account for the cyclic variations in fluid chemistry that have been observed to have a tidal frequency (Sugisaki, 1981; Sugisaki and Sugiura, 1986): earth-tide driven fluctuations in hydraulic heads of subsurface aquifers have been widely documented (Roeloffs, 1987; Roeloffs and Bredehoeft, 1985) . Such fluctuations could generate significant chemical variations in a spring discharge for an indefinite period of time without causing a progressive, irreversible, change in the up-flow zone as would be required by the models invoking pore collapse or micro-fracturing.

In summary, the currently available laboratory and field results suggest that changing stress levels prior to seismic events can, possibly by dilating existing joints and fractures in the subsurface, open (or re-open) fluid pathways that were formerly sealed. Such changes in permeabilities (and hydraulic heads) could account for not only the changes in water chemistry, but also the very erratic changes observed in temperatures, flow rates, and water levels. The trends in these hydro-geochemical variations would be strictly the result of the existing hydrologic conditions in the aquifers being mixed. Hence, a well tapping two artesian aquifers may form (or enhance) a connection to the one having a higher hydraulic head and begin to fountain whereas other higher level springs or wells being fed by the same one may cease flow entirely. Similarly, changes in ion concentrations will be controlled by the compositions of

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the fluids in the two (or more) aquifers being mixed in a spring or well-bore conduit. Thus, even though the precursory changes in fluid chemistry may appear to be random in a given area, they may well be responding to a regional stress change that occurs prior to an impending seismic event.

SUMMARY

The mechanisms that seem to be best able to account for the reported precursory geochemical and hydrologic anomalies are the Increased Reactive Surface Model and the Aquifer Breaching/Fluid Mixing Model. The former mechanism can account for many of the gas concentration anomalies that have been reported but only for modest increases in ion concentrations. The latter seems best able to explain the major changes in ion and isotope chemistry as well as water levels, flow rates, and temperatures that have been observed. Models invoking ultrasonic vibration, pressure sensitive solubilities, and pore volume collapse seem less able to explain the range of chemical anomalies that have been detected in precursor monitoring programs.

APPLICATION TO EARTHQUAKE FORECASTING

Presuming that the above mechanisms can adequately explain the observed precursory phenomena, the major remaining question is how, or whether, the geochemical variations can be used to forecast an impending earthquake. Although several researchers have claimed that the timing, rise time, magnitude, or distribution pattern of the geochemical precursory anomalies can be used to predict the location, timing, and magnitude of an impending earthquake, none of these claims have held up to close scrutiny or, more importantly, to a test of reproducibility. The reasons for this can be attributed in part to the mechanisms responsible for the anomalies and in part to the protocols followed in the geochemical monitoring programs and interpretation of the data gathered.

An examination of the Aquifer Breach/Fluid Mixing Model clearly shows why any attempt to extract earthquake magnitude information from the magnitude of the geochemical anomalies is destined to failure. Even though the extent of fluid interchange or mixing that may accompany stress induced changes in aquifer head or flow rate could be a function of the magnitude of the stress changes, it is evident that the impact of this mixing on a water source being monitored will be strictly a function of the respective heads and fluid compositions of the aquifers: extensive mixing of fluids having identical chemical compositions would produce no detectable geochemical anomaly but only minor minor changes in the flow of two markedly different aquifers would produce a substantial anomaly. Hence, it is virtually impossible to predict, a priori, the magnitude of a geochemical anomaly that may correspond to a given magnitude earthquake without having complete knowledge of the chemical and hydrologic

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characteristics of all aquifers that might interact with the water source being monitored.

Similarly, the distribution of the groundwater chemistry anomalies around an impending epicenter is likely to be a complex function of the sensitivity of the water source being monitored to the local hydrologic system and, more importantly, how the precursory changes in the stress/strain field are transmitted to or through the hydrologic system. Models of the stress strain distribution around an impending epicenter (Dobrovolsky, 1979) have shown complex distributions of compressional and dilational stress in the far field that, when imposed on a hydraulic system, could have effects that will be very difficult to interpret without an excessive monitoring effort. Of even greater concern, however, is the fact that the
theoretical models have only been able to evaluate the
stress/strain fields around faults, or asperities on faults, as though the surrounding area is an isotropic homogeneous half-space. It clearly is not; subsidiary faults, elevation changes, and variable shapes and orientations of asperity volumes, are all likely to have a profound effect on the stress distribution patterns around an impending hypocenter. Hence the expectation of a uniform, predictable, distribution of precursory hydrologic and fluid chemistry anomalies around an asperity volume is unrealistic given the complexities of multiply faulted terrain.

Ground-gas anomalies in the concentrations of hydrogen and radon are not believed to be sensitive to the same hydrologic variables that groundwater chemistry variations are. None-the-less, they are subject to the same uncertainties of distribution and transmission of the stress/strain field as the hydrologic anomalies and to other interferences such as changes in meteorological conditions or variations in soil permeability. Thus, even though ground gases variations may not have the inherent unpredictability of the response of groundwater chemistry anomalies, their interpretation is not straightforward.

In spite of these difficulties, however, I believe that useful data can be obtained from precursory geochemical variations that are associated with seismic activity. In order to best extract this information, it will be necessary to institute a long term, continuous (rather than an intermittent) monitoring program of several geochemical and hydrologic variables that should include both ion chemistry as well as dissolved gas and ground gas
chemistry. The regional choice of a monitoring station should
be based upon geophysical considerations of where significant seismic events are likely to occur. However, site specific location of monitoring stations should maximize sensitivity to seismic stress changes and minimize sensitivity to surface
effects. Analysis of the data generated should include
statistical pattern recognition and Fourier, or other spectral,

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analysis techniques such as that applied by Shapiro et al. (1985). In order to alleviate the currently unknown manner in which the geochemical anomalies are likely to be distributed around an impending epicenter, the monitoring program would have to be broadly based and, as the sensitivity, or lack there-of, of various sites is determined, the distribution of monitoring stations should be modified accordingly.

A monitoring program similar to that outlined above should be able to detect both geologically anomalous conditions - such as those associated with creep events or pre-seismic stress changes as well as non-geologic interferences that may impede the identification of pre-seismic conditions. The identification of geologically anomalous conditions would make it possible to then forecast the probability of a seismic event in a given region; the absolute prediction of the precise location and time of a seismic event is not likely to be possible until many of the current uncertainties related to the geologic structure of a given area, and its probable response to a changing stress field, can be resolved.

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GAS-GEOCHEMICAL APPROACHES TO EARTHQUAKE PREDICTION

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Abstract

Concentrations of a wide range of terrestrial gases in ground water and soil air have commonly been found to be anomalously high along active faults, suggesting that the faults may be paths of least resistance for the outgassing process of the "solid" earth. Temporal gas-concentration changes with dura-tions of a few hours to many months have been discerned before many large and some smaller earthquakes at stations mostly located along active faults at epicentral distances of as much as several earthquake-source dimensions where

the associated coseismic strain changes are estimated to be as small as 8 10 . This result suggests that the earthquakes and the associated anomalies are both incidental results of some small but broad-scale episodic strain changes in the crust. Such strain changes may be amplified at the earthquake and anomaly sites and, together with sufficient pre-existing stresses, may reach some critical levels (above half fracture strengths) for generation of the earthquakes and anomalies. Significant gas-concentration changes have also been observed at times of other known crustal disturbances, such as underground explosions, ground-water pumping, and earth tide, and from rock samples in laboratory experiments. These results suggest several possible mechanisms for the deformation-related gas-emission changes that involve movement of crustal fluids of non-uniform chemical composition or enhanced water/rock reaction at newly created rock surfaces. However, gas-concentration changes may also be caused by various non-tectonic environmental changes, which must be recognized in the search of true earthquake precursors.

Introduction

Scientific measurements for earthquake prediction have been made exten-sively since the mid-1960's in many seismically active countries such as USSR, China, Japan, and USA. The reported geophysical and geochemical data often show temporal changes interpreted to be premonitory to some large or moderate earthquakes. Some of the reported anomalies are considered by other investi-gators to be marginal in their tectonic significance, for similar changes may be caused by other concurrent environmental variations such as rainfall or groundwater pumping. In a previous paper (King, 1986) I gave an overview of gas-geochemical studies carried out in various seismic regions of the world and discussed possible effects of various environmental changes. In the present paper, I show a collection of representative gas-geochemical data sets published in the literature and suggest a physical basis for the observed features in the data, which still need to be substantiated by further studies.

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Spatial Anomalies and Active Faults

Concentrations of a wide variety of terrestrially generated gases (e.g., radon, helium, hydrogen, mercury, and carbon dioxide) contained in ground water and soil air have been found to be anomalously high along active faults, suggesting that the faults may be paths of least resistance for gases in the "solid" earth to escape to the atmosphere. Fig. 1 shows, for example, the result of Israel and Bjbrnsson (1967) who measured radon and thoron concentra-tions in soil air at 1-m depth along transects perpendicular to the strikes of several faults near Aachen. They attributed the radon anomalies that were accompanied by thoron anomalies to enrichment of parent nuclides, and the other radon anomalies to upward migration of radon.

Irwin and Barnes (1980) showed that the worldwide distributions of springs in which the ground-water HCO^ content exceeded 1,000 ppm generally coincided with major seismic belts (Fig. 2). Fig. 3 shows helium anomalies along active faults, especially at intersections of faults, observed by Yanitshiy et al. (1975), who measured helium concentration in ground water at depths of 30-50 m in north Kazakhstan. Similar anomalies were also observed for several other gases (e.g., C02, Rn, and Hg).

Temporal Variations Associated with Known Crustal Disturbances

A number of field experiments have been conducted in which the gas con-centrations in. ground water and soil air were found to change in response to known crustal strain changes caused by such events as underground explosions, ground-water pumping, and earth tide. Fig. 4 shows soil-air radon anomalies observed by Wollenberg et al. (1977) at five monitoring sites near the nuclear explosion Esrom in the Nevada Test Site. The anomaly amplitudes decreased with increasing distance from ground zero, and were not related to barometric pressure variations recorded nearby.

Fig. 5 shows the result of Sugisaki (1981), who observed a correlation between tidal strain and the He/Ar concentration ratio in gas bubbles from a mineral spring in Japan. Similar results were also observed for other gases

at several other "sensitive" sites, suggesting that terrestrial gas emissions 8 at such sites may be responsive to crustal strain changes as low as 10

Earthquake-Related Gas Changes

Gas-concentration changes have been reportedly observed before many large
and some smaller earthquakes in various parts of the world. Some of these changes might be coincidental, caused by non-tectonic environmental changes, such as rainfall and ground-water pumping, or by instrumental or human error. Others are probably truly related to earthquakes, and they appear to show some general features in common: (1) They tend to occur mostly along

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active faults, especially at the intersections and bends of faults, at epicen-tral distances up to several times the corresponding earthquake-source dimensions. (2) The durations range from a few hours to many months. The occurrence, or the increasing occurrences, of the short-term (spike-like) anomalies appears to be useful to predict the time of earthquakes (perhaps within a few days to a few weeks). (3) The spatial extent and the duration of the intermediate-term anomalies generally appear to increase with, and thus may be useful for predicting, earthquake magnitude. (4) The amplitude of the anomalies does not appear to show consistent correlation with either earth-quake magnitude or epicentral distance (in contrast to the case of explosion shown in Fig. 4). Following are some representative examples of earthquake-related gas changes, which are deemed relatively reliable, because they were recorded in multiple components, by independent methods or observers, at multiple sites, and can be compared with a reasonably long set of background data.

Radon Anomalies Before the Tangs nan Earthquake

Radon content of ground water has been monitored at many wells in north-
ern China shortly after the magnitude 7.2 Xingtai earthquake in 1966 (see King, 1985, for a review of radon studies in China). Fig. 6 shows the loca-tions of the magnitude 7.8 Tangshan earthquake in 1976 and monitoring stations. Most of the stations that recorded the supposedly intermediate-term and/or short-term anomalies (increases, which are sometimes difficult to recognize, in Figs. 7 and 9) were located along major or secondary fault zones up to an epicentral distance of about 500 km (Figs. 8 and 10). The intermedi-ate-term anomalies (gradual increases beginning in late 1973) may be related to both the Haicheng (1975, magnitude 7.3) and Tangshan earthquakes on account of their proximity in time (1.5 years) and space (350 km). One station (Guanzhuang) was in operation long enough to cover the magnitude 7.4 Bohai earthquake in 1969 also; the associated anomaly, which peaked after the earth-quake, is shorter in duration but comparable in amplitude. The anomaly amplitudes are small, ranging from 10 to 100% and do not depend significantly on epicentral distance (Figs. 8 and 10). Short-term anomalies (spike-like increases mostly, Fig. 9) were recorded in increasing numbers at the wells a few days to weeks before the Tangshan main shock and several strong after-shocks (Fig. 11). However, very few of them were recorded in the immediate epicentral areas.

The above-mentioned radon anomalies typify those observed elsewhere in
the world. Fig. 12 shows a plot of anomaly amplitude vs. epicentral distance for a worldwide data set compiled by Hauksson (1981). The areal extents of anomaly occurrences are also unexpectedly large, and increase with magnitudes of the associated earthquakes. The threshold of the related strain changes was estimated to be ~10 , comparable with tidal strain changes.

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Gas-Geochemical Anomalies in Southern California During 1979

Gas-geochemical measurements were made extensively in southern California
by several research groups. Fig. 13 shows helium and radon concentrations recorded by Chung (1985) and colleagues at Arrowhead hot spring on the San Andreas fault. These concentrations varied synchronously and were anomalous at the time of several significant local earthquakes including the magnitude 4.8 Big Bear earthquake in 1979, but not the magnitude 4.1 earthquake in 1983. Similar anomalies were recorded for methane, nitrogen, electric conductivity, and temperature, and they were all attributed to admixing of gas-rich water from a deep source with the regular shallow water within the spring. No such anomaly was recorded, however, at this group's other 15 monitoring sites. Teng and colleagues (see Teng and Sun, 1986) also recorded significant anoma-lies in 1979 at three of their 14 ground-water radon monitoring stations located within 20 km of the Big Bear earthquake (Fig. 14). So did Shapiro et al. (1985) at some of their monitoring sites along a frontal fault of the Transverse Ranges (Fig. 15). Shapiro et al. considered the anomaly recorded at Kresge between June 1979 and early 1982 to be tectonic in origin, possibly as the result of a regional strain episode, which may be responsible also for the increased number of moderate earthquakes in southern California, including the magnitude 6.6 Imperial Valley earthquake on October 15, 1979 about 290 km away from their sensitive stations.

Geochemical Anomalies During 1980 in Central California

King et al. (1981) and O'Neil and King (1981) recorded some chemical changes in ground water at two wells along the San Andreas fault near San Juan Bautista, California (Fig. 16) at the beginning of a sequence of earthquakes in 1980; the largest of the earthquakes has a magnitude of 4.8 and epicentral distances of several kilometers. Fig. 17 shows the result of in situ measure-ments of water level, temperature, salinity/electric conductivity, and pH at a 30-m deep artesian well (Mission Farm Campground). The water level had gener-ally been rising since 2 years before the earthquake; it reached the ground surface (the well became self-flowing) at about the same time as a magnitude 4.0 earthquake that occurred 25 km away, or about 2 months before the magni-tude 4.8 earthquake at an epicentral distance of 6 km. Simultaneously, the water temperature became more steady, and the salinity/conductivity increased by many standard deviations from the background level. Similar results were obtained from chemical analyses of water samples (taken from the well at the same times as the in situ measurement) in the laboratory, especially for contents of Ca++, Mg++, F~, Cl", NO^, SO^~, and HCO^ (Fig. 18) and stable isotopes 6D and 6 O (Fig. 19a). The isotopic contrast is shown more clearly on a plot of <So versus 6^ 8O (Fig. 19b). Such changes were attributed to admixing of water from another aquifer that was normally not tapped by the well, possibly through cracks created in an intervenient aquiclude during a tectonic strain episode. Similar but generally opposite changes were observed at another well (San Francis Retreat, SFR in Fig. 16) about 80 m deep and about 3 km southeast of the first (Figs. 20-22). However, no significant

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changes were recorded at a third well (Cienega Winery, CW in Fig. 16) somewhat farther away from the epicenter area. Measurements at the former two wells were interrupted shortly after the anomalies because the wells were then used for continuous radon monitoring that requires pumping.

Gas-Geochemical Anomalies Before the 1984 W. Nagano Earthquake

Gas-geochemical anomalies were recorded by three groups of researchers (Sano et al., 1986; Sugisaki and Sugiura, 1986; Katoh et al., 1986) at several stations near the magnitude 6.8 earthquake on September 14, 1984 in the west-ern Nagano Prefecture of central Honshu, Japan. Sugisaki and Sugiura (1986) observed conspicuous gas anomalies at a fumarole and three springs about 1-3 months before the earthquake. Fig. 23 shows the He/Ar, N? /Ar, and CH./Ar concentration ratios in bubble gases from a mineral spring (Byakko) on an active fault about 50 km from the epicenter. They attributed the anomalies to changes in the emission rate of some deep-seated gas due to earthquake-related changes in pore pressure. Sano et al. (1986) observed some post-earthquake increases (compared with values obtained in November 1981) of He/ He ratios in bubble gases from several hot springs located less than 10 km from a fault that was possibly formed at the time of the earthquake. They suggested that the earthquake was triggered by an upward migration of some He-rich fluids associated with magma intrusion beneath the source region. Katoh et al. (1986) observed anomalous increases of radon concentration in soil gas (super-posed on some apparently seasonal changes) at several monitoring stations located on three active faults about 25 to 100 km away from the epicenter, beginning about 1 to 2 years before the earthquake (Fig. 24). These anomalies coincided with coda-wave duration anomalies observed by Sato (this volume). They are similar in shape to soil-gas radon anomalies recorded elsewhere previously (see, e.g., King, 1978).

Possible Mechanisms for Earthquake-Related Gas-Geochemical Anomalies

Several possible mechanisms have been postulated by various investigators
to explain the earthquake-related gas anomalies, most involving tectonically induced movement of crustal fluids of different concentrations or enhanced water/rock reaction at newly created rock surfaces. They include increased upward flow of deep-seated fluids to the monitored aquifer or holes, gas-rich pore fluids being squeezed out of rock matrix into the aquifer, mixing of water from another aquifer through tectonically created cracks in the inter-venient aquiclude, increased water/rock interactions, and increased gas emana-tion from newly created cracks in the rocks to the pore fluids. Some of these possibilities have been tested in laboratory experiments. For example, Holub and Brady (1981) monitored radon emanation and microcracking activity in a uranium-bearing granite sample under uniaxial loading. They observed an initial decrease in emanation possibly associated with closure of pre-existing cracks in the rock and then significant increases in both the emanation and microcracking activity at a load of about one-half of the ultimate strength

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(Fig. 25). Such increases were observed repeatedly until larger increases occurred at time of fracture. Kita et al. (1982) crushed granite and quartz samples under moist conditions at temperature of 25°-270°C, and found increased H2 release with temperature up to about 200°C; at higher tempera-tures the release decreased with temperature (see Fig. 26 for granite). They attributed the H_ generation to chemical reaction between water and Si and Si-O radicals on the fresh surfaces of the crushed rocks.

King (1978) invoked a tectontically-induced vertical flow of crustal gases to explain soil-gas radon anomalies recorded along the San Andreas fault in central California. As shown schematically in Fig. 27, radon concentration generally decreases slowly with height in near-surface atmosphere, and increases rapidly (by 3 orders of magnitude) with depth in near-surface soil gas, approaching a maximum at a depth of several meters. Because of the high subsurface concentration gradient, a small upward/downward flow can signifi-cantly increase/decrease the radon value monitored at a fixed shallow depth of about 1/2 m. King considered a one-dimensional model in which the soil is treated as a homogeneous porous half-space with uniform radon production rate. On the assumption that radon migrates by molecular diffusion governed by Fick's law and by flow governed by Darcy's law and that the radon concen-tration is negligibly small at the surface, it was shown that the steady-state concentration C is a function of depth -z as follows (e.g., Clements, 1974):

C =

where

Y =

2(eAD) 1/2

<j) is radon production rate, X is its decay constant, e is soil porosity, D is molecular diffusion coefficient of radon in the soil, and v is apparent soil

gas flow velocity (volume of flow per unit time per unit geometric area). 4 Fig. 28 shows this result for five different flow velocities (0, ±3 x 10 , ±10 cm s~) and for some material constants appropriate for typical dry 4 ** 1 soils. It is evident that a small upward/downward gas flow (3 x 10 cm s)
can significantly increase/decrease the subsurface radon concentration at shallow depths (by a factor of 2 at a depth of 0.7 m). Note that much larger increases in upward gas velocity do not produce much larger increases in radon concentration; this result may explain why amplitudes of observed radon anomalies are not correlated significantly with earthquake magnitudes and epicentral distances.

King and Slater (1978) compared the soil-gas radon data recorded at two sites along the Calaveras fault in Hollister, California, with crustal strain data recorded on a two-color laser geodetic meter in the same area; they found periods of higher radon emanation to be coincided (with slight delay) with periods of crustal compression (in north-south direction), as expected from King's model, with the cross-correlation coefficient being 0.67 (Fig. 29). The variations in both data sets may be partly seasonal in nature. Neverthe-less, the high radon values in 1979 may be partly related to the magnitude 5.9 Coyote Lake earthquake on August 6, 1979 about 27 km away on the same fault. 286

Similar models may explain earthquake-related anomalies in other soil-gas components (He, 1^, Hg, etc.) that have similar concentration-versus-depth profiles. Concentration gradients may also exist in wall rocks of aquifers, which may be considered as consisting of networks of "flow pores" (Fig. 30; see Norton and Knapp, 1977) that transport meteoric waters from recharge areas to the monitored springs or wells. Since ground water in the flow pores usually has shorter residence time in the ground and less reaction with rocks, it contains less dissolved terrestrial gas (and other chemical substances) than the fluids in the "diffusion" and "residual" pores in the wall rocks (Fig. 30). Thus the tectonically induced flow from the normally diffusive and isolated pores to the flow pores is expected to introduce waters of anoma-lously high gas and ion content.

The same models may explain the results of some large-scale pumping tests conducted in China, where the radon contents in the pumped wells tended to increase, whereas those in the surrounding observation wells to decrease (see King, 1985).

Strain Field Responsible for Earthquakes and Anomalies

The general observation that gas-geochemical (and other) anomalies are widely spread in space but confined mostly to active fault zones suggests that they (and the associated earthquakes) are possibly incidental results of some broad-scale episodic strain changes in the crust (King, 1986). Such changes can be produced by a number of processes, including magma intrusion and epi-sodic fault-creep events below seismogenic depths. The amplitudes of change _ Q need not be very large (£,10), but may be greatly amplified along pre-existing fault zones and especially at their intersections and bends. The local strain increases at such places, when combined with sufficient pre-existing strains, may reach critical levels (above half fracture strengths) to generate the observed tectonic and physical anomalies (as defined by Ishibashi, this volume) and earthquakes. The anomalies may disappear before or after the earthquakes for a variety of reasons, including stress relaxation associated with the earthquakes, healing of the fault zone, and exhaustion of gas supply (in the case of gas anomalies). The concept of multiple-point concentration of a broad-scale episodic strain change may also explain why earthquakes and the various anomalies sometimes occur in distinct yet appar-ently related clusters in space and time along the same or different faults, why it seems difficult to predict earthquake location with reasonable preci-sion, why there is a lack of detectable seismic-wave velocity anomalies (the dilatant zones may be too thin compared with seismic wave lengths and too widely distributed), and why there is a lack of observed one-to-one corre-spondence between earthquake and anomaly occurrences.

One piece of evidence supporting the possibility of strain amplification along a fault zone is coseismic steps recorded on creepmeters along the San Andreas fault (King et al., 1977). If interpreted as strain steps (see example in Fig. 31), their amplitudes are about 2 orders of magnitude larger

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than expected for the strain field of the associated earthquake, based on a dislocation model on the assumption of a homogeneous crust (Fig. 32).

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Figure 1. Concentration of radon and thoron at 1-m depth in soil air over faults near Aachen. The concentration was measured in transects at right angle to the strike of the faults, which had previously been located by geo-logical methods. A section of the bedrock under each transect is schemati-cally indicated. (After ISRAEL and BJORNSSON, 1967.)

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Seismic zone CO 2 locality

Figure 2. Worldwide distribution of seismic zones (s'haded) and localities of springs with high C02 discharges (HCO~1 content > 1000 ppm). (After IRWIN and BARNES, 1980.)

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Figure 3. Anomalous high helium concentration in groundwater at depths of 30-50 m along consolidated and slightly permeable fault zones (dashed lines) (A), and along active fault zones (B), especially at intersections of faults in north Kazakhstan. The helium content in 10~ ml/S. is <1 in white areas, 1-12.5 in sparsely dotted areas, and 12.5-125 in the densely dotted areas. (After YANITSKIY et al., 1975.)

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Figure 4. Temporal variations of soil-gas radon concentrations monitored at five shallow holes (0.7 m deep) near underground nuclear explosion Esrom in the Nevada Test Site. Holes 1-5 are, respectively, 0.2, 1.3, 2.7, 3.8, and 4.6 km from ground zero. Mean values and standard deviations from the pre-Esrom values at each site are indicated by error bars. Barometric pres-sure data at the NTS are shown at the bottom for comparison. (Modified from WOLLENBERG et al., 1977.)

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Figure 5. Comparison of temporal variations of He/Ar concentration ratio in gas bubbles in a mineral spring at Byakko Spa, Japan (circles) with theoreti-cal tidal strain (areal dilation; dotted curve), atmospheric pressure (solid curve), and temperature (dashed curve) for two time periods. (After SUGISAKI, 1981.)

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Figure 6. Locations of monitoring stations that showed intermediate-term ground-water radon-concentration changes (circles with horizontal stripes), short-term changes (circles with vertical stripes), and no significant change (small circles) before the 1976 Tangshan earthquake (solid circle) in northern China. Thick solid and dashed lines indicate observed and inferred major faults, respectively. (After YING et al., 1978.)

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Figure 7. Time series of ground-water radon concentration (monthly average values) recorded at sensitive stations in northern China. Arrows indicate the times of 1975 Haicheng and 1976 Tangshan earthquakes. (After YING et al., 1978.)

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Figure 8. Amplitude of intermediate-term ground-water radon anomalies (relative to background level in monthly average values) as a function of epicentral distance from the 1976 Tangshan earthquake. (Based on data in YING et al., 1981.)

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Figure 10. Amplitude of short-term ground-water radon anomalies (relative to background level in daily readings) as a function of epicentral distance before the 1976 Tangshan earthquake. (Based on data in YING et al., 1981.)

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Figure 11. Frequency of occurrence of short-term radon anomalies recorded at water wells in northern China. Indicated are occurrences of Tangshan earth-quake and larger aftershocks (magnitude > 6.0, circles). (After YANG et al., 1982.)

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Figure 13. Time series of helium and radon concentrations in water samples taken monthly from the Arrowhead hot spring on the San Andreas fault in southern California. Vertical lines indicate the times of several local earthquakes with labeled magnitude. (After CHUNG, 1985.)

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Figure 14. Time series of radon concentrations in water samples taken from three USC stations along the San Andreas fault within 20 km of the magnitude 4.8 Big Bear earthquake (indicated by arrows) in southern California. (After TENG and SUN, 1986.)

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Figure 15. Time series of ground-water radon concentrations recorded at four Caltech stations in southern California. (After SHAPIRO et al., 1985.)

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Figure 16. Location of water wells (circles) and earthquakes (stars) of magnitude 4.0 or larger that occurred during the study period of KING et al. (1981).

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Figure 17. Time series of water level, temperature, salinity, electric con-ductivity, and pH at the MFC well. Horizontal lines indicate pre-anomaly mean values ± one standard deviation. Vertical lines indicate occurrence of earth-quakes of magnitude > 4.0; labeled are magnitude and epicentral distance in km. (After KING et al., 1981.)

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Figure 20. Time series of water level, temperature, salinity, electric conductivity, and pH at the SFR well. Symbols as in Fig. 16. (After KING et al., 1981.)

Figure 21. Time series of ion concentrations (mg/L) in water samples from the

SFR well. NOl and NO! are in mg/L N. (After KING et al., 1981.) 2 -J

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Figure 24. Time series of radon concentration in soil gas (TD: a-particle track density in weekly-exposed cellulose nitrate film; MA: moving average) at three sites located on the Median Tectonic Line, the Matsushiro fault, and the Atera fault, which are, respectively, 100, 100, and 25 km away from the 1984 Western Nagano prefecture earthquake. (After KATOH et al., 1986.)

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Figure 25. Radon emanation (upper curve) from a uranium-bearing granitic rock under uniaxial stress (lower curves). (After HOLUB and BRADY, 1981.)

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Figure 26. Amount of H2 release from crushed granitic rocks at various temperatures and in two different atmospheres. (After KITA et al., 1982.)

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RADON CONCENTRATION

Figure 27. Schematic diagram of radon concentration in atmosphere and soil air near the ground surface as a function of elevation and depth concentration in undisturbed soil air at great depth; C
surface. (After JUNG, 1963.)

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Figure 28. Radon concentration as a function of depth in a homogeneous, porous half-space with uniform radon production rate for five different verti-cal gas flow velocities. (Modified from KING, 1978.)

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Figure 30. Schematic representation of flow pores ((f>p), diffusion pores
and residual pores (<f> R) in rock. Fluid usually flows through the flow pores but not the diffusion or residual pores, unless the rock is significantly stressed. (After NORTON and KNAPP, 1977.)

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Figure 31. Spatial distribution of coseismic shear strain steps from creep-meter recordings along the San Andreas fault for the magnitude 5.0 Melendy earthquake in 1972. Dots represent data from creepmeters spanning the fault; circle, not spanning. The lower half of the figure shows the hypocenter of the earthquake (cross) and aftershock area (shaded area) projected on the fault plane. (After KING et al., 1977.)

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Geochemical precursors: Short-term and intermediate-term anomalies

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Abstract

Continuous monitoring of the radon concentration of groundwater in Japan was effective in one case in detecting the short-term anomaly of a nearby earthquake. With the exception of the 1978 Izu-Oshima-kinkai earthquake (M7.0), however, no other abnormal change has been noted. This may partially be due to difficulty in detecting insignificant precursory signals from observation data which ordinarily fluctuate. The temporal variations in the radon concentration changes are site-depen-dent and vary significantly, affected by various factors origi-nating in activities near the ground surface. In order to detect precursory phenomena of an earthquake, better understanding of the mechanisms controlling radon emission at every observation site is necessary. In addition to the present radon observation, He/ He ratio measurement may be useful in detecting pre-seismic signs of inland-type earthquakes.

1. Introduction

Research efforts aimed at detecting precursors of earth-quakes have been under way in China, the USSR, the USA, Japan, and other countries. A numbers of overviews of geochemical precursors have been published (Raleigh et al., 1977; Wakita, 1978a, 1982; Teng, 1980; Zhang and Fu, 1981; King, 1985; Barsu-kov et al., 1985). Geochemical observation is still at the stage of accumulating data and searching for parameters effec-tive in detecting precursory phenomena of an earthquake. The number of clear precursors is too small to indicate plausible features of the phenomena. At the same time, however, it must be stated that successful prediction of several destructive earthquakes which occurred in China was made based on anomalous changes in geochemical data. Precursory changes are usually recognized as abnormal changes in the records. On the basis of experiences over 10 years in Japan, we see the most urgent problems as (a) clarifying the difference between normal and abnormal changes in observing data and (b) building a model that causes precursory changes.

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2. Definition of abnormal changes

Because there are no reasonable explanations for the causes of precursory changes, any kind of change that differs from the normal trend can be regarded as abnormal in the loosest sence. This may lead to an erroneous understanding of precursors, however. Abnormal changes must be identified based on an understanding of the characteristics of the observed data. In general, precursory signals are weak and their detec-tion is not easy. Background fluctuations in the observed records caused by changes in meteorological conditions such as temperature, atmospheric pressure, heavy precipitation, and tidal motions, and by changes due to human activities are rather large and show irregular patterns. If the magnitude of the precursory change of an earthquake is smaller than the magnitude of such interferences, its detection will ultimately be impossible. Since a precursory change may also be detected through other data with an amplified mode, comparison of geo-chemical data with various other kinds of observations is extremely helpful.
As a typical example, various precursory changes of the Tangshan earthquake (China) are shown in Figure 1. The radon concentration changes themselves, shown as (2) and (3), are quite obscure. In contrast, changes in leveling (1), magnetic field (4) and resistivity (5) are rather distinctive. These changes seem to be contemporaneous to the occurrence of two earthquakes. If we examine just the radon data, we may not be able to recognize their peculiarities. However, comparing the patterns and occurrence times of the radon changes with other data including those for seismic events, we can see to some extent the precursory features. Radon concentrations (2) and
(3) decreased in association with the occurrences of the two earthquakes. Marked fluctuations in radon concentration (3) became calm after the earthquakes. These kinds of changes are representatives of the reported precursors.
In contrast, a sudden eruption of groundwater from wells and/or springs is significant and is easy to recognize as a precursor. Even though the observation is clear, however, there remain many questions. Matsuda and Sibano (1965) report-ed a good example of abnormal changes prior to a M6.1 earth-quake which occurred in Shizuoka, in the Tokai region (Japan), on April 20, 1965. One of the natural gas wells located in the epicentral region gushed water several times before the earth-quake, beginning in the early part of the year (Figure 2). Water eruptions occurred once in January, three times in Feb-ruary, four times in March, and once in April. The last eruption occurred April 1, 19 days before the 1965 Shizuoka earthquake. These eruptions can be considered short-term pre-cursors of the earthquake.
Prior to these water eruptions, a peculiar occurrence of water flowing was observed on June 7, 1964, about 10 months before the earthquake. Nine days later, the Niigata earthquake

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(M7.5) occurred, with an epicentral distance of about 380 km. Pointing out the particular geotectonic situations of the epi-center of the Niigata earthquake and the well, both located at the edges of the active tectonic zone of the Fossa Magna, Mogi (1981) suggested that the water flow might be related to the destructive earthquake.
Other than on these occasions, water flow of the No. 11 well (whose depth is 231.5 ra) had not been noted since its operation began in 1948. In other words, no water flow was observed when three other large earthquakes (magnitudes of 7.5, 7.0 and 6.9) occurred during the period between 1948 and 1964, within areas less than 400 km away from the well. How should we interpret the water flow of June 7? Was this an intermediate-term precursor of the M6.1 earthquake or a short-term anomaly of a M7.5 earthquake, or unrelated to these earthquakes? Another big problem is that precursory changes are not observed in all observation sites in a given area. Even at the same site, as in the above instance, only one particular well was sensitive, while many others were insensitive. The appearance of precursors is quite capricious. Even though one earthquake has clear precursors, the next earthquake occurring in the same region may have no precursor. An example is the cases of the 1978 Izu-Oshima-kinkai earthquake (M7.0) and the 1980 Izu-Hanto-toho-oki earthquake (M6.7), which occurred in almost the same region about one and a half years apart. Al-though many obvious precursors were observed for the 1978 Izu-Oshima-kinkai earthquake, as will be described in a later section, practically no meaningful change was observed for the 1980 Izu-Hanto-toho-oki earthquake.
According to Mogi (1986), the appearance and magnitude of precursory signals are attributed to differences in the crustal structure and stress history of the region. The results of a rock fracture experiment showed that occurrences of acoustic emission events of rock differ between the first loading and the second loading. The activity of the first loading is markedly higher than that of the second loading. If the inter-val between the two loadings lengthens, the difference disappears. The experiment suggested that the stress history remains in rock for a certain period. This means that the magnitudes of precursory phenomena will be small in an area which is being stressed by the occurrence of an earthquake in the near past.
To further complicate the matter, abnormal changes are not always earthquake-related: even after changes which have all the characteristics of pre-seismic events, there may be no earthquake. This may in some cases be attributable to an inherent resistance of the region to earthquakes.
Thus, in addition to actual observation, a physical model
for the causes of precursors is definitely needed. Such a model may also help in detecting the precursors. Similarly, studies on the mechanism controlling radon emission changes are also important. The formation of new cracks, closure of

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fissures in rocks, migration of groundwater from different sources, and changes in the artesian layer, all due to an earthquake in preparation, are considered probable causes of radon emission changes. Since variation patterns of observation data are site-dependent, information on the degassing features of radon at every observation site is necessary to distinguish meaningful changes related to an earthquake from those derived from background fluctuations.

3. A short-term precursor of the M7.9 Kanto earthquake

Case studies of clear precursors of past earthquakes are helpful not only for detecting faint signals of an earthquake but also for building a physical model of precursors.
One of the most remarkable precursors is water flow of hot springs and wells. Nakamura (1925) reported a clear short-term precursor to the Great Kanto earthquake (M7.9), which occurred on September 1, 1923. The Ooyu geyser, a famous geyser in Atami hot spring, had been active until the end of 1890 but had become gradually inactive, and no geyser eruption had been observed since December 20, 1922. The geyser, however, became active again in early May 1923. The first recovery of activity was noted on May 8 and 9, when eruptions continued for 19 and 17 minutes, respectively (Figure 3). Successive activities that commenced at the end of June continued to the time of the earthquake. During the period, the erupting activity fluctuated significantly. The most significant eruption was observed on August 31 , one day before the earthquake. It lasted for 40 min and struck the local people as quite abnormal.
A few minutes after the main shock on September 1, the geyser erupted suddenly, and significant water flowing con-tinued for about 10 days. After that, the flowing activity gradually decreased, and ceased in March 1924, six months after the earthquake.
As shown in Figure 3, it is difficult to recognize the intermediate-term precursor. Was the geyser eruption in May a precursor? Alternatively, the cessation in activity after December 1922 might have been a kind of intermediate-term precursor.
Because of its history, this geyser is a useful sensor for predicting a future Kanto earthquake. At the present time, it is erupting on a small scale. The decrease in geyser activity is partially attributed to recent large-scale commercial development in the area, including drilling of deep wells.

4. Short-term precursor

The only clear example of a short-term precursor observed
in Japan via geochemical investigations is in the case of the 1978 Izu-Oshima-kinkai earthquake (Wakita et al., 1980; Wakita,

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1981), when abnormal changes in the radon concentration of groundwater were recorded at an observation well. The radon concentration at the site began to decrease in the middle of December 1977 and fluctuated rapidly thereafter. It remained at a lower level for a while, reaching the minimum value on January 8, six days before the main shock. Then the concentra-tion level began to increase suddenly and returned to the pre-vious higher level two or three days before the earthquake. A magnitude 7.0 earthquake occurred on January 14, 1978. An increase in the radon concentration caused by vibrations due to the main shock and aftershocks was also noted. The patterns and occurrence times of the abnormal radon changes well resem-ble those of water temperature and water level in two different artesian wells located nearby, and of strain measured by a borehole strainmeter (Figure 4). The observation sites are spread over the Izu Peninsula. The similarity implies that these changes are caused by a single tectonic force in the area before the M7.0 earthquake.
The observed precursory changes of the radon concentration differ markedly from the variation patterns of regular years. The annual variation pattern over eight years, between 1978 and 1985, is superimposed on the year of the earthquake in Figure
5. The heavy dotted line represents the average variation over eight years calculated from the normalized pattern of each year; the difference is readily apparent.
Anomalous changes beginning in the middle of October 1977
are significant. There seems to be a sequence in the pattern of precursors: normal > abnormal » normal > earthquake. Recovery to a normal level prior to an earthquake is an impor-tant precursor.
It is worth mentioning that the abnormal period of about
two to three months for the M7.0 earthquake is comparable to those of about two months observed for the M6.1 Shizuoka earth-quake (Figure 2) and the M7.9 Great Kanto earthquake (Figure 3). When these changes are taken into consideration, the duration of the short-term precursors is about 2 to 3 months, regardless of the magnitude of the earthquake.
As seen in Figure 5, a rapid change occurred a few days before the earthquake. This change is very similar to the imminent-term radon concentration changes observed at the Guza station in China (Figure 6), where a spike-like change was occasionally observed a few days before several large earth-quakes. The location of the station at the intersection of three major fault systems may act to increase sensitivity in reflecting precursory stress change in the area.

5. Intermediate-term precursors

The definition of the occurrence time and duration of the intermediate-term precursor is flexible. In this paper, the time is tentatively considered to stretch from half a year to a

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year prior to an earthquake. The detection of the intermediate-term precursor is much more difficult than that of the short-term precursor. The expected patterns may become very broad due to the longer continuation of abnormal changes. These k'inds of variations may easily be hidden in long-term varia-tions such as seasonal change. In order to detect an abnormal change, continued reliable observation over a sufficiently long period that includes the abnormal period and the occurrence of the earthquake is necessary. The emerging question is how we can correlate the observed changes with successively occurring earthquakes. There are many choices in selecting one earth-quake from among others with varying times, places and magni-tudes. So far no persuasive intermediate-term precursor has been reported.
Figure 7 shows the long-term record of the radon concent-ration changes observed at SKE site, where the background fluctuations were least pronounced among our observation sites. Earthquakes with magnitudes greater than 5 and hypocentral distances of less than 100 km and those larger than M6.9 which occurred in more distant areas are plotted in the figure. It is difficult to find any correlation between radon changes and these earthquakes, except for the 1978 Izu-Oshima-kinkai earth-quake. The magnitude of intermediate-term precursors is expect-ed to be extremely small compared to background fluctuations, even though they existed. The adoption of a new parameter originating from a different source will be helpful in increasing knowledge of geochemical precursors. The defining characteristics of the intermediate-term precursor are least affected by changes in environmental factors.

6. Helium isotope ratio

Emission of helium is also applicable to estimating changes in the stress in the Earth's crust, given helium's ideal characteristics of mobility, chemical inertness, and low abundance in the atmosphere. In general, determination of the absolute concentration of gas is affected by changes in other gas components: addition or removal of one gas will change the concentrations of the remaining gases considerably. The isotope ratio, however, is generally independent of these effects. Furthermore, use of the helium isotope ratio has an additional merit. Helium has two stable isotopes, ^He and He. A large portion of J He in terrestrial samples is primordial helium derived from the Earth's mantle. The other isotope, He, is produced mainly by radioactive decay of U and Th in the crust. Thus, a high He/ He ratio implies a closer relation to magmat-ic activity.
The 3 He/ He ratios of terrestrial samples vary in a wide range reflecting their tectonic environments: greater than 2 x 10"^ for hot spots, 1.2 x 10~ 5 for oceanic ridges, less than 9 x 10~~ 5 for subduction zones, and less than 1 x 10~ 7 for conti-

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nental regions. The former three categories are closely related to magmatic activities; the latter is the least related. Atmospheric He has the ratio of 1.4 x 10~ . The helium in the samples is composed of three different He sources: atmospheric, magmatic and radiogenic (crustal). The He/ He ratio will change in accordance with varying mixing ratios of these components, possibly caused by tectonic force. This systematics can be applied to monitor the stress change in the crust.
Figure 8 shows the geographical distribution of He/ He ratios of gases (Sano and Wakita, 1985) and that of microseis-micity in the Japanese islands. The closed and open circles imply 3 He/ He ratios greater than and smaller than twice the atmospheric value, respectively. The closed circles are mostly concentrated in the backarc region. The distribution of the He/ He ratios coincides well with that of shallow earthquakes in the archipelago, suggesting a closer relationship between these two observations.
Most inland-type earthquakes occur in the backarc region
of the Japanese islands. Since some of their occurrences are closely related to magmatic activities, changes in the 3 He/ He ratio are applicable to detect precursory changes. Since variations in the He/ He ratio are least influenced by changes in environmental conditions, the ratio is suitable to use in detecting abnormal changes in a longer scale.
The He/ He ratio system can be applied to investigate the triggering effects of an earthquake. The 1965-66 Matsushiro earthquake swarms, for example, are thought to have been caused by a diapiric uprise of a magma as a result of He/ He measure-ments (Wakita et al., 1978b). The causes of the 1984 western Nagano earthquake (M6.8) (Sano et al., 1986) and of continuing Wakayama seismic swarm activity (Wakita et al., 1986) are also discussed in terms of changes in the 3 He/ He ratio. In the above interpretation, fluids expelled from a cooling magma increase the pore pressure and act to trigger the occurrence of an earthquake in a preexisting tectonic stress field. The introduction of magmatic helium due to upward migration may chancre the observed He/ He ratio. Periodic measurements of the He/ He ratio will useful to increase knowledge on intermediate-term and long-term precursors.

7. Conclusions

Several conclusions are drawn from this study.
1) The duration of short-term precursors is 2 - 3 months, irrespective of the magnitude of the earthquake.
2) Imminent changes may be observed a few days before an earthquake. The changes are expected to be very sharp, with step-like patterns.
3) Due to lack of sufficient observation data as well as
of occurrences of large earthquakes, the recognition of mid-

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term precursors is rather difficult.
4) Our most urgent need is to build a physical model of
the appearance of precursory changes.
5) A means of measuring H e/He ratio sensitive to magmatic movements will be useful for inland-type earthquakes in the future.

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Time sequence of water eruption

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Niigata Eq.
June 16
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A = 380km

Shizuoka Eq.
April 20
M 6.1
A~0

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May June July Aug. Sept. Oct. Nov. Dec. Jan. Feb. Mar. Apr.
1964 1965

Figure 2. Time sequence of water eruption observed at a natural gas well in Shimizu in the Tokai district. The bottom figure is redrawn based on the original by Matsuda and Sibano (1965). Occurrences of water eruptions and of five large earth-quakes with epicentral distances of less than 400 km, including the 1964 Niigata earthquake and the 1965 Shizuoka earthquake, are plotted (upper part). No other water eruption was noted since operation of the well began in 1948. Numbers on bars are the dates of water eruptions (lower part).

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Atami Hot spring
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Kanto Eq.
M 7.9

Sept. 1, 1923

March

1923

April May June July Aug. Sept.

Figure 3. Time sequence of water eruption of Ooyu Geyser at Atami hot spring. The duration of eruptions varied significant-ly prior to the Great Kanto earthquake. Epicentral distance was about 30 km. The figure is drawn based on the data in a report by Nakamura (1925). A similar attempt was made by Hiraga (1981). After the earthquake, water eruption continued for about 10 days. The record of these observations is thought to be reliable, since it was made by policemen on duty.

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lzu~O»hlma-klnkal earthquake
14 January 1076 (M7.0)

SEP. OCT. NOV. DEC. JAN. FEB.

1977 1978

Figure 4. Precursory changes of the 1978 Izu-Oshima-kinkai earthquake (Wakita, 1984).(1) Radon concentration change observed at an artesian well (the epicentral distance: d = 25 km); (2) Water temperature variations observed at a 500 m well (d = 30 km); (3) Water level variations at a 500 m well (d = 30 km); (4) Strain changes measured by a borehole strainmeter at the tip of the peninsula (d = 50 km).

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Izu-Oshima-kinkai Eq.

M 7.0

January 14, 1978

SEPT. OCT. NOV. DEC. JAN. FEB. MAR. APR

Figure 5. Precursory radon changes of the 1978 Izu-Oshima-kinkai earthquake observed at SKE well. A heavy dotted line shows the average annual variation calculated from data for eight years. Note the significant decrease and rapid fluctua-tions beginning in the middle of October. A step-like change that occurred a few days before the earthquake is also meaning-ful. The rapid increase after the earthquake was attributed to vibration effects due to the main shock and aftershocks.

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Figure 8. Geographical distributions of He/ He ratios of emerging gases (upper part) measured by Sano and Wakita (1985), and of microseismicity (lower part) in the Japanese islands. Closed and open circles in the upper figure imply He/ He ratios greater than and smaller than twice the atmospheric value. The VF lines are the volcanic front. A seismicity map of shallow earthquakes (0-15 km) during the period between July 1 983 and June 1 984 is from data compiled at the Earthquake Prediction Center, Earthquake Research Institute, University of Tokyo. Due to lack of seismic observation, some part of Kyushu and Hokkaido districts remain blank.

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SUMMARY OF SESSION ON HYDROLOGY AND GEOCHEMISTRY

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Roeloffs gave a very balanced and objective review of the possibilities and problems of using instrumented water wells as strain meters. While many examples of hydrologic precursors have been reported, none is unequivocal. Because water levels and pressure head changes depend on barometric pressure, rainfall, and nearby pumping, it is always difficult to eliminate nontectonic effects. Water levels appear to be a very sensitive strain indicator within a limited frequency band. That band is limited at the important low frequency end (about 10" 6 Hz, or 1/10 days), by fluid flow within the reservoir. For higher frequencies water wells may have sensitivity and signal/noise ratios comparable to borehole dilatometers. At Parkfield, the response of pressure changes to volume strain has been estimated both from tides and coseismic strains, and the agreement is quite satisfactory for most wells. Thus many wells may be reasonably well calibrated by their tidal response.

Broedehoft and Rojstaczer gave an exemplary discussion of the response of a water well to volume strain. A well is a complex system responding to atmospheric pressure variations as well as fluid pressure changes throughout the aquifer. Because there is a time lag between a pressure head disturbance anywhere in the aquifer and a change in water level, the well and aquifer must be viewed as a system with frequency dependent amplitude and phase response. But the amplitude and phase are not independent, and the transmissivity of the well can be estimated from its phase spectrum alone. Both water level changes and downhole pressure head (for shut in wells) or volume changes must be known to calibrate a well. In most cases theoretical volume tides must be substituted for direct downhole pressure or volume measurements, introducing phase errors of up to several degrees. Each well has its own character, depending on its diameter, the aquifer it taps, and how open it is to the atmosphere.

Oki and Hiraga took a more empirical approach, listing several examples of apparent hydrological precursors observed by amateur observers (the Catfish Club) in Japan. The best example seemed to precede the 1978 West Off Izu-Oshima earthquake (M 7.0). Water level anomalies of up to 100 mm were observed in wells at epicentral distances from 60 to 220 km, about one month before the earthquake. There were also water temperature anomalies at 36 and 40 km, and borehole strain (dilatometer) anomalies at Ajiro (35 km) and

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Irozaki (40km). While the occurrence of so many anomalies over such a large area is hard to explain by anything besides tectonic strain, it is also true that some fairly close sites showed no preseismic anomaly, and that the reported anomalies were quite different in detail. Oki and Hiraga used the observed sensitivity of volume strain to barometric pressure (10 /Pa) and the hydrostatic pressure gradient (10 Pa/mm) to obtain a strain sensitivity of 10 9 mm (that is, 1 m of water height for 10~ 6 strain). They did not mention whether the response of the wells to tidal strains was consistent with the above sensitivity.

Generally speaking, water wells can be extremely sensitive to strain within their frequency band, and they have a tremendous cost advantage over more sophisticated borehole dilatometers. However, their response is poor at periods longer than a few weeks, and their response at shorter periods may be complicated. While excellent progress has been made in understanding wells, much more need doing to apply these new ideas if water wells are to be used responsibly in earthquake prediction.

Wakita, Nakamura, and Sano gave a rather complete review of the search for geochemical precursors in Japan, with additional comments concerning work in China. They found only one example of a clearly anomalous radon change in ground water preceding an earthquake: that was the 1978 West Off Izu-Oshima earthquake. As reported above, this earthquake was also preceded by variations in ground water flow and volumetric strain at several sites around the Izu Peninsula in the month or so before the event. The radon anomaly is said to be similar to the ground water and strain anomalies. However, the most dramatic radon variation occurs six days before the event, and I saw no evidence of ground water or strain variations at that time. The authors point out that radon and other geochemical indicators respond to atmospheric variations, rainfall, and many other causes, not all of them understood or even known. Thus the noise level is high for geochemical precursors, and recognizing anomalies is somewhat of an art.

King summarized his global review of geochemical methods, now published (J. Geophys. Res., 91, 12269-12281, 1986). He gave many examples, and I cannot effectively discuss individual cases here. He emphasized that many anomalies occurred hundreds of km from the earthquakes they preceded, while closer sites experienced no geochemical anomalies. He explained this paradox by invoking "sensitive" stations. Many of the anomalies are observed at sites within fault zones, or at intersections of fault zones, etc. These sites might produce a greater chemical perturbation for a given stress than other sites because of strain concentration, reservoir tapping, or other special properties.

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Thomas summarized nicely the types of geochemical techniques used in earthquake prediction studies and the theories advanced to explain hypothesized precursors. Observations include both the absolute concentrations and the abundance ratios of metallic cations, radioisotopes, and dissolved gases, sampled from soils, flowing ground water, and deep wells. Possible mechanisms for precursory changes include
(1) solubility changes from ultrasonic vibrations, (2) solubility changes from pressure variations, (3) stress induced pore collapse, (4) creation of fresh reactive surfaces by stress variations, and (5) stress induced aquifer breaching. While none of these mechanisms adequately explains all of the claimed precursors, Thomas favored (5) as being most capable of producing large geochemical changes.

Personally I find almost all of the reported hydrological and geochemical precursors to be totally unconvincing as tectonic phenomenon. No doubt the observations reflect real changes in hydrology or geochemistry, or both, but the connection to tectonic stress changes is dubious. Few of the proposed anomalies are observed at more than one site, few are accompanied by comparable or larger coseismic anomalies, few demonstrate a believable relation between anomaly size and epicentral distance, and none satisfies all three of the above criteria. Carefully monitored and calibrated water wells can be used as narrow band strain meters, but in the reported cases of preseismic anomalies the wells were not calibrated nor did they behave in a spatially coherent manner. The observed geochemical variations bear almost no relationship to the location of the associated earthquake, they give many false alarms, and they do not accompany many earthquakes. There is no satisfactory theory to explain why the minute stress changes that might precede earthquakes would cause such large geochemical effects as those reported. On the other hand, the chemistry of terrestrial fluids is known to depend on many nontectonic effects, and it is this reporter's opinion that most of the geochemical anomalies are unrelated to earthquakes.

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Resistivity and Magnetic Variations as Earthquake Precursors:
A Search for Possible Causes

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Resistivity Variations

In the heyday of optimism about earthquake prediction based on the dilantancy-diffusion concept there were always nagging points that stuck out. Resistivity anomalies reported in the U.S.S.R. and used to support the dilatancy diffusion concept (Schulz et al, 1973) showed resistivity decreases that would be expected if dilantancy was occurring (provided the partial saturation was not too dramatic), but a rather symmetric recovery with no apparent coseismic resistivity change. The data base was not nearly good enough for any convincing argument to be made against the concept, and it is still conceivable that a dilantancy induced resistivity decrease would not be erased by the earthquake stress relief. We still lack an adequate data base, but the best data available leads us to consider the possibility of different processes. The resistivity data collected in China in the region around Tangshan is the most remarkable such data set that this author is aware of and leaves one with many ideas to think about (Ma, 1982; Qion, 1981). The data set is made even more dramatic by the knowledge that this disastrous earthquake was not predicted.
The Chinese collect resistivity data using active sources and rather
small spreads, (~ 1 km). The data is recorded manually from measurements that are repeated several times during each 24 hour period. Figure 1 shows the long term (5 years) resistivity decrease reported by the resistivity stations in the region. The fact that the data contours so well over such a large area given the rather shallow depth of investigation of these measurements is truly remarkable. Figure 2 shows daily averages of the recorded measurements for a period of weeks just prior to and after the earthquake. These data have a very different character than the long term trends as they are much more erratic both in sign as well as in magnitude.
Figure 3 shows graphs of the long term behavior for two stations for which long term data after the event was also available. This data as well as that shown in figure 2 emphasize the general lack of coseismic effects and leave one with a definite feeling that the earthquake caused a jump in the rate of change of resistivity but not a jump in resistivity.
When looking back at other reported resistivity anomalies one can see
this as a general pattern. The resistivity anomalies are usually decreases of resistivity and coseismic effects are minor if at all present.
The lack of a coseismic resistivity signature is perhaps an important
sign. Many of the earlier concepts that were put forth to explain precursors involved stress changes, but the resistivity data may perhaps be better explained as the effect of stress rather than stress changes. This author was always a bit concerned about the resistivity changes reported, since he also used to think in terms of stress changes and using laboratory data on the effect of stress or rock electrical properties was led to not expect very large anomalies. Table 1 is a compilation and extrapolation (and some guessing) of expected resistivity changes due to stress changes (T. Madden, 1978) as determined from the laboratory data. These results are based on very short term measurements which one might describe as the elastic response.

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Table 1
Resistivity Stress Sensitivity percentage resistivity change per bar stress change

Porosity in % 1 3 10 30

Depth in km 0 .5 .4 .1 .03

1 .3 .2 .1 .03

3 .1 .1 .05 .03

10 .05 .05 .03 .03

It is now well recognized that stress can cause slow changes such as
stress corrosion and pressure solution. Pressure solution is an important component of the consolidation seen in sedimentary rocks, but this is a one way process and would not be well suited to explaining the repeated behavior of earthquake related phenomena. Stress corrosion, however, is a state of dynamic balance and could well fit into the earthquake proccess. Materials under nonhydrostatic stress are not in a state of thermodynamic equilibrium and as argued by Gibbs surface matter from material under nonhydrostatic stress would be expected to dissolve in order to redeposit on surfaces of the same material under hydrostatic stress. Crack tips with their high stresses are just such surfaces. But Gibbs also points out that surfaces are not thermodynamically stable in situations where they can be reassimilated into the solid. Crack tips with their high concave curvatures are just such surfaces. Both processes involve reaction rates and can be expected to occur slowly at low temperatures. They are also contradictory processes and the net direction that the processes take will depend on the state of stress. High stresses should lead to stress corrosion and low stresses will favor crack healing.
Laboratory studies of stress corrosion (Scholz, 1972; Rutter et al,
1978; Atkinson, 1986) show rather steep dependencies on stress and these studies would not lead one to expect much of a stress activated change at the stresses and temperatures that are likely to be found in the upper crust. It is often the case, however, that different processes are involved and that another process with a lower stress dependency could dominate at lower stresses. Such low stress corrosion studies still need to be done in order to clarify this issue.
Even if the hypothesis that the resistivity decreases seen before
earthquakes are due to stress corrosion and the resistivity increases seen after earthquakes are due to crack healing is correct, the value of resistivity measurements for earthquake prediction would still be dependent on some relationship between stress corrosion and earthquake phenomena. The issue was well put by Scholz (Scholz et al, 1973) when he expressed the idea that precursors were signals of the inevitability of a coming earthquake. At this state of our knowledge such statements are somewhat pulled out of thin air, but since part of our quest here is to generate ideas for consideration we will offer up a bit of thin air.
Many of the so-called asperities that lock up fault zones are the
significant deviations of the fault surfaces from planer geometry. Unlocking such a barrier must involved some sort of plasticizing of the surrounding media (This is a concept borrowed from Geoff King.) which for brittle material can be achieved by creating a hierarchy of interconnected cracks of different sizes. The development of this plastic like condition is a stress corrosion

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process triggered by the stress load exceeding some critical level and this development is also responsible for the electrical resistivity decreases.

Differential Magnetic Variations
Differential magnetic variation measurements are technically more
difficult to make because precise differencing of magnetic field measurements between different sites is needed which requires either good communication links or very accurate timing. An extensive array exists in California
and measurements are also made in Japan and on several volcanic islands. The source of the anomalies are changes in the magnetic susceptibility or remanent magnetization of rocks in the vicinity of the recording site. To date these anomalies are interpreted as the effects of stress changes called the piezomagnetic effect (Stacey et al, 1972) and this suffers much of the same problem as the resistivity variations. Anomalies associated with volcanic activity do seem to be reasonably well explained as a piezomagnetic effect (Johnston et al, 1969, Stuart et al, 1975) where large stress changes over relatively short time spans can occur. The California magnetometer array does record variations that are real and some (San Juan Bautista, November 1974) even occurred near to and shortly before an actual earthquake, but it seems doubtful that the data can be easily explained as a piezomagnetic effect. This author knows little about long term effects on magnetization but there is no reason not to suspect that stress corrosion-like factors may also be involved in modifying magnetic properties. This is an area that is probably worthy of investigation, but it may take more dramatic examples of earthquake-related magnetic anomalies to push such investigations.

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Fig. 1 Long (3-4 years) and short (3-4 months) term resistivity variations associated with Tangshan earthquake of July 28, 1976 (M 7.8). Contours represent long-term variations.
Profiles show short-term measurements on arrays with typical separations of 1 km. Time
scale for short-term variations shown at left. (Ma Zongjin et al., 1972; Oian Jiadong, 1981)

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Fig. 3 Long-term near surface resistivity variations, Tangshan Region (monthly averages)

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GEOELECTRICITY IN EARTHQUAKE PREDICTION toward a model and method of prediction

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ABSTRACT

This article will review geoelectric methods that have been applied
to earthquake prediction research in China, Greece, Japan, USA and the USSR. The resistivity response of stressed rock will be discussed for both low and high strain levels. It will be shown (1) that resistivity

- 7 8 measurements can monitor strains as low as 10" to 10 and (2) that some of the electrical precursory phenomena may be related to pressure solution rate phenomena. Furthermore, it will be emphasized that the electrical measurements must sense predominantly the near-surface zone in order to obtain high sensitivity levels.

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INTRODUCTION

In this report I will attempt to show that shallow (1-3 km) resis-tivity measurements can be used as a sensitive low-cost continuous stress/strain monitoring system. Relevant measurement techniques, electronic instrumentation and optimal field siting will also be discussed.
In attempting to establish a theory or methodology for earthquake prediction, based on the resistivity/stress/strain concept, I will rely on four main classes of information, as follows:
(1) Computing the theoretical stress/strain and thence the resis-tivity/strain distributions associated with a fault, leads to the conclusion that most of the resistivity changes are in the assumed focal region and also in the near surface zone. The surface zone
is a good target because of its wide lateral extent.
(2) Experimental data from Japan, China and the US will be used to show that there is an amplification between resistivity and strain. The amplification which can be as large as 10 3 to 101*, is very non-linear, increasing rapidly at low strains. I will also provide a plausible explanation for the effect. Moreover, it will be shown that the amplification factor is greatest in low porosity rocks.
The near-surface zone represents the region with the highest sensitivity to this enhancing non-linear resistivity/strain behavior of rocks.
(3) I will briefly review virtually all attempts at earthquake predic-tion using geoelectric methods. This will demonstrate that the successful experiments have been those which monitored resistivity
in the near-surface zone.
(4) There are now strong indications that some resistivity variations are due directly to stress via a pressure solution mechanism. It
is suggested that this also takes place preferentially in the near surface.

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Unfortunately, from the measurement standpoint, the near-surface environment is very noisy. Noise in this case comes mainly from temperature and water table (rainfall) fluctuations. To circumvent or reduce these noise problems, one of two approaches is suggested:
(i) monitor resistivity (from the surface) at many depths and interpret the data in such a manner as to reduce or remove the
surface noise,
(ii) monitor resistivity in or between shallow (*\» 100 m) boreholes. Furthermore, siting of the array is of critical importance. Preferable sites are those in low porosity rocks and not in high porosity materials such as sandstones or highly fractured rocks.
Taken together a model emerges, which indicates that it is indeed reasonable and possible to monitor strain in a cost-effective manner, by using appropriate near-surface resistivity measurements and interpreta-tion methods.

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ATTEMPTS TO USE ELECTRICAL METHODS IN EARTHQUAKE PREDICTION

Many attempts have been made to measure changes in the electrical signature prior to and during earthquakes. Both passive and active, near-surface and deep-probing measurements have been implemented. Some of these experiments are briefly described below.

Experiments in the USA
Cancelled Tellurics (MIT)
Madden (1984) has operated, for more than 10 years, a passive experiment which utilizes long dipoles for the measurement of the telluric field. The experiments are located at Hollister and Palmdale, California. Dipoles are typically of the order of 10's of km. At least three telluric dipole signals are "cancelled" against each other to produce a residual that is highly sensitive to resistivity variations. The resistivity changes must occur differentially under the array. If the entire area undergoes a resistivity change of equal magnitude, then the cancellation would not be offset. The method claims sensitivity to better than 0.1%, over periods of years. To date, Madden has not seen any resistivity changes that are conclusively correlated with earthquakes.
Shallow Resistivity (MIT)
Fitterman and Madden (1977) reported on a surface resistivity measurement at Melendy Ranch, California which ran from mid-1972 to Jan-1974. The experiment was designed to investigate possible correlations between resistivity and surface creep events. The experiment used an approximate Schlumberger array where the current electrode spacing was 200 m and the voltage electrode spacings 60 m and 23 m. The transmitter and receiver were co-linear and in close proximity to the surface exposure of the fault. The creep meter was placed within the resistivity array. Short-term sensitivities to 0.005% were achieved by temperature controlling the electronic instrumentation. No correlation with slip events as large as 3 mm was observed. Fitterman and Madden (1977) attributed this negative result to the fact that the subsurface was mainly composed of clays. The argument is that continued

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slip occurring in clay, produces aligned slip planes with little associated resistivity change.
Intermediate Depth Dipole-Dipole Resistivity (UCLA)
A 3.3 km dipole-dipole experiment was conducted near Palmdale, California during 1977-1978 (Searls et al., 1978). The system used a 20 kW source and 900 m dipoles. Accuracy on the resistivity measurement was ±1%. After U years of operation the only significant resistivity change (6%) seemed to be associated with rainfall and the concomitant water table variations.
Deep Dipole-Dipole Resistivity (U.C. Berkeley)
Mazzella and Morrison (1974) and Morrison et al. (1979) reported on
a deep dipole-dipole experiment across the San Andreas fault, just south of Hoi lister, California. The dipoles were of 1.5 km length and the dipole-dipole separations were approximately 10 km and 15 km apart. The power was generated from a 80 kW source, sending 200 A with a 10s period into the ground. Many cycles of the received signal were stacked. Accuracies of ±2% were achieved. Approximately three months after these measurements began a 24% precursory resistivity change seemed to be associated with a magnitude 3.9 earthquake which occurred within the resistivity array (Mazzella and Morrison, 1974). In 1977 a magnitude 4.0 earthquake occurred in virtually the same position. There was no associated resistivity change. The reconciliation is not clear (Morrison et al., 1979). However, it is possible that the initial 24% anomaly was indeed an experimental artifact.
Shallow Borehole-Surface Resistivity (NOAA)
Bufe (1973) and Bufe et al. (1973) established a near surface active resistivity experiment close to the U.C. Berkeley experiment and the MIT creep monitoring resistivity system at Melendy Ranch, California. The electrodes were distributed in an approximate Wenner array with an inter-electrode spacing of 200 m. The current electrodes were in boreholes approximately 100 m deep. The potential electrodes were buried to 3 m. The system accuracy was 0.1%. Bufe (1973) noted clear correlations between earthquake induced strain changes and resistivity variations (Figure 1). Figure 1 shows one surprising result. The strain change is on the order of 10" whereas the

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_3 accompanying resistivity variation is only of the order of 10 . There 4 appears to be an "amplification" of 10 in the resistivity case. This observation agrees well with those in Japan, as discussed below. Long term fluctuations of approximately 10% were noted. These were apparently produced by temperature and rainfall variations affecting principally the near surface potential electrodes and electronic instrumentation.

Experiments Outside the USA Deep Dipole-Dipole Resistivity (USSR)
Barsukov (1970), Barsukov and Sovokin (1973) discuss a large scale dipole-dipole resistivity measurement in the Garm region of the USSR. The system transmitted high current (100 A) pulses which were received approximately 10 km away. Many conductivity changes of approximately 10% were recorded. These changes were apparently precursory and there seemed to be some correlation between the magnitude and duration of the conductivity changes and the magnitude of the earthquake. Some of these changes could possibly be associated with the effects of temperature and rainfall on the electrodes and instrumentation. The current status of this experiment is unknown.
Near-Surface Telluric Field Studies (Greece)
Since early 1981 Varotosos and Alexopoulos, from the University of Athens in Greece, have been conducting telluric studies related to earthquake prediction. A pair of brass electrodes 50 m to 100 m apart are used to receive the telluric signal at frequencies less than 0.3 Hz. There are now nearly 20 stations operating in Greece. The signals are telemetered back to Athens in real time. The results of this on-going study were reported by Vorotsos and Alexopoulos (1984a, 1984b).
They have noted precursory behavior associated with many earthquakes. Moreover, they have been able to deconvolve the precursory signatures to such an extent that they can estimate the location to approximately 50 km and the magnitude to less than 0.5 unit accuracy. Earthquakes with epicenters as much as 500 km away have been predicted. Precursor times range from 6 to 115 hours. In order to validate the technique, these researchers have been logging telegrams of predictions

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before the seismic events. For the period January 19, 1983 to October 21, 1983 and for earthquake magnitude ^ 5, 23 seismic events occurred and 21 of these were predicted. During the period October 1982 to October 1983, 24 events with seismic magnitude >_ 4.3 occurred within the network. All events were predicted.
In summary, this means that the claimed success rate of prediction
is better than 70%! This is an astonishing result! No detailed model of the process was proposed. However, it was suggested that stressed rocks generate a large transient electric field. This author knows of no viable process to explain their observations and views their results with great scepticism.
Small-Scale Resistivity Variometer (Japan)
The Earthquake Research Institute of the University of Tokyo operates a resistivity variometer consisting of a high accuracy auto-matic 67 Hz a.c. bridge which monitors the resistivity across three 1 m to 2 m electrode spreads situated within a cave. The system has been in operation since 1968 and is described in a number of papers by Yamazaki and Rikitake (Rikitake and Yamazaki, 1969a, 1969b; Yamazaki, 1965, 1966, 1967, 1968, 1974, 1975). Many co-seismic and precursory resistivity changes have been recorded with this instrument system for earthquakes up to distances of nearly 1000 km and magnitudes between 5 and 7.

- 7 -8 Strains as low as 10 - 10 have been recorded even though the
- 4 variometer has a sensitivity of approximately 10 in Ap/p. Note that
- 7 -8 at 10 to 10 strain, the induced tidal loading can be recorded and
this is in fact used as the means of calibrating the variometer.
The extra "gain" of 10 3 to 101* comes from a non-linear effect in which the relative change in p is greater at smaller strains. This behavior is shown in figure 2. We will offer a physical explanation of this effect in the next section. Note that the "gain" is rock-type dependent, being greatest for the Lapilli tuff which comes from the Aburatsubo site where the variometer is installed.
Shallow Resistivity (China)
There are a number of groups in China, but principally at the Lanzhou Earthquake Research Institute and the Institute of Geophysics, State Seismological Bureau in Beijing, that have been systematically measuring resistivity as a means of predicting earthquakes since 1967. There are now more than 100 resistivity stations in China.

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Typically, measurements have been done using current electrode spacings of 1 km with a Wenner array. Usually two measurements are taken in perpendicular directions. The depth of penetration is there-fore approximately in the 200 m to 300 m range. The results of some of these studies have been reported by Qian J. et al. (1979), Qian J. (1985) and Qian F. et al . (1983). They have observed changes in conductivity, often with precursory times of years, prior to most of the major (M ^ 7) shallow (h ^ 30 km) earthquakes within 200 km to 300 km of the array. One reassuring aspect of their data is that superimposed on the long term precursory variation one can observe the yearly seasonal changes. Some of this data for the Tangshan earthquake is shown in figure 3. Figure 4 shows the regional variation of resistivity. The epicenter of the main shock was located close to the maximum resistivity change of -4.4%. Note in closing that the Chinese have also conducted deep dipole-dipole resistivity which has shown no precursory effects. Summary
The above review is fairly complete although there have been a few other attempts, principally using tellurics and magnetotellurics, that I have not discussed. They have not been particularly successful, and at any rate, not very pertinent to the present article.
The experiments reviewed can be summarized as follows:
Successful
China (Resistivity) (Shallow)
USA (Resistivity) (Shallow)
Japan (Resistivity) (Shallow)

USA (Tellurics) Deep Greece (Tellurics) Shallow Russia (Dipole-Dipole) Deep China (Dipole-Dipole) Deep Unsuccessful
USA (Dipole-Dipole) Intermediate USA (Dipole-Dipole) Deep
What this summary clearly shows is that active resistivity measurements that sample the shallow subsurface, on average, seems to be the most

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successful methods. The following section attempts to show, from a rock physics viewpoint, why near-surface probing possesses the highest sensitivity.

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STRESS-STRAIN-RESISTIVITY BEHAVIOR

Two regimes will be considered for the stress-strain-resistivity response of rocks subjected to the tectonic forces responsible for earthquakes. The first case considers the resistivity changes associated with small stresses or strains. The second category looks at the resistivity variation in the vicinity of a theoretical strike-slip fault as the rocks approach shear failure.
Volume Strain-Induced Resistivity Variations
Consider Archie's law,

A ~m /1 \ p = P Q <f> (1)
where p is the resistivity of the rock saturated with solution of resistivity p . <J> is the porosity. Take the index m to be 2. In differential form this equation can be written, approximately, as,
Ap m AV ~P " " * ~V

where AV/V is the fractional change in the rock volume. It is assumed that all the volume change occurs in the pore-space. Co-seismic

- 5 earthquake strains are usually 10 or smaller. Earth tides induce
- 7 -8 strains in the 10 to 10 range. Long term strain rates along the San
Andreas are usually in the range 0.1 to ly strain/year, (see for example Rikitake, 1976). If resistivity is to be used as a stress/strain monitor of tidal and co-seismic effects, Ap/p could possibly range from 10" 2 to 10" 7. Figure 5 shows a plot of (Ap/p)/(AV/V) against AV/V, for $ between 0.2% and 20%, which clearly indicates that the resistivity changes are "amplified" by factors ranging from about 10 to 10 3 depending on the porosity. Note that the "amplification" or "gain" is independent of strain.
In practice (Ap/p)/(AV/V) vs (AV/V) is not independent of strain.
The data (figure 2) of Yamazaki (1965) clearly shows this effect. Data from Brace et al. (1965) can be plotted in a similar form, as shown in figure 6 for the case of Westerly Granite. The amplification factor reaches approximately 500. Note that the strains in this experiment are

. 3 approximately 10 . The amplification is therefore expected to be -5 larger than 500 for strains as small as 10 or less. The non-linear

amplification is caused by the preferential closure of cracks in the low

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strain regime. For a fully saturated rock compressive stress will increase the resistivity whereas for partially saturated rocks the conductivity will increase with stress. There are many other examples of this amplification effect, both from laboratory and field studies, documented in the Chinese literature. Figure 7, taken from Qian Fuye et al. (1986) is a compilation based on earthquake data. Again we see amplifications of the order 10 3 with a few as high as 101*. As one might expect this non-linear amplification is rock-type dependent, being largest in low porosity rocks possessing a large crack porosity. An approximate way of accounting for the above behavior is to write equation (2) as,
Ap _ m f AV / 3 \
j. I l^ i

where the porosity <j> is actually a non-linear function of strain <J>(AV/V). It is also probable that "m" varies, being close to two (2) at high strains and decreasing towards one (1) at very low strain levels. However, sufficient data is not available to adequately test this idea. Before closing this section it may be worthwhile to point out two further things. The amplification or increased sensitivity at low stress levels applies to virtually all physical properties such as velocity, porosity, resistivity, permeability, etc. The degree of sensitivity varies with each property. Permeability will show very high sensitivities but is extremely difficult to measure in-situ. Resistivity shows high sensitivities and the measurement is straightforward. The second point is to note that the sensitive behavior occurs at low (< 1 kb) stress levels. Clearly the stress-strain-resistivity sensitivity is therefore greatest in the near surface zone.
Shear Strain-Induced Resistivity Variations During the late 1960's and early 1970's a number of papers were published on the process of dilatancy (Brace et al., 1965; Brace et al., 1966; Brace and Orange, 1968a; Brace, 1975; Nur, 1972; Scholz et al., 1973; Scholz and Kranz, 1974). The model suggests a "dilatant" stage which preceeds the actual earthquake rupture process. During the dilatant stage, the porosity is increased with associated changes in seismic velocities, electrical resistivities, fluid flow, etc.

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I will consider a strike-slip fault because of its relevance toward earthquake prediction along the San Andreas fault in California. Fitterman (1976) has already tackled this problem and the following treatment draws heavily on his presentation.
Figure 8 represents an ideal strike-slip fault diagrammatically. Prior to an earthquake, no slippage takes place above the depth d on the fault plane, below this, to infinity, slippage with negligible friction is allowed. With time, the strain increases to some critical value, and is then released in the form of an earthquake which ruptures the earth about d. Quite often, this rupturing extends to the surface. Because the model is two dimensional, it is also assumed that the fault has infinite length in the y direction.
The seismic displacement of such a system may be represented mathematically by the semi-empirical relationship, (eg. Weertman and Weertman, 1969 or Savage, 1972):

Uy (x, Z) = sf ^an- 1 £ + tan -l .£_} (4)

where x and z are the horizontal and vertical distances from (0,0) as shown in Figure 8.
Where faults extend to the surface in California the relative displacement between the two sides is of the order of 2.5 cm/year. Assume that for a locked fault as in Figure 8, the relative displacement b is identical to this value multiplied by the time interval considered. Fitterman (1976) used such a screw-type dislocation model with resistivity vs. shear-failure data to estimate the % change in resistivity for cross-sections perpendicular to strike-slip faults. Figure 9 is taken from his study. Most of the resistivity variations (0.5 to 1%) occur in the top 1 km for the specific models choosen. Fitterman choose a model with the fault locked to 15 km depth. If instead a 3 km depth is used, the resistivity changes are then approximately 4% and increase as the assumed hypocenter approaches the surface. Figure 10 shows conceptually that the regions with the largest resistivity changes are the surface zone and the area in the immediate vicinity of the assumed hypocenter.

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Summary
Amplifications of 10 3 to 104 between resistivity and strain seem to
be evident from many types of data. Resistivity-strain shows the highest sensitivity in the near-surface zone whether low or high strain levels are considered. Measurements which sense the top 1 to 3 km for conductivity variations will perhaps exhibit the highest sensitivity and offer the greatest chance of successfully obtaining resistivity percursory signatures.
The reason that there is an increased resistivity change close to the surface is due to the fact that the resistivity response to shear stresses is a strong non-linear function of depth. The model of Fitterman (1976) makes a number of simplifying assumptions. However, the general conclusions are probably correct. We may therefore conclude that resistivity changes of the order of 1.0 to 0.1% may occur in the top 1 to 2 km. Clearly the region in the vicinity of the assumed hypocenter will show intrinsically larger resistivity changes, however because of its depth this is a much more difficult "target". Moreover, any surface method will sample such a substantial volume in attempting to sense. the hypocentral region that the large intrinsic resistivity change at the focus will be diluted substantially.

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TOWARD A MODEL

Let us make an attempt to construct a model based on the field data
and our understanding of the stress-strain-resistivity behavior, as discussed above. Most of the data relevant to this problem comes from China.
Consider again figure 3. The data shows on average a 1% to 5% decrease in resistivity prior to the Tangshan earthquake, with a time constant of 2-3 years. The post-earthquake resistivity increases at a very low rate. In many instances, and also with other data-sets, the resistivity seems to level off at a lower value than before the earthquake, indicating some permanent pore-space deformation or reorientation. Qian F. (1983) points out that such behavior is similar to that seen in the stick-slip experiments of Wang et al. (1978). Figure 11, taken from Wang et al. (1978) shows the resistivity continually decreasing, while also responding instantaneously to the shear stress steps. However, because of the long time constant associated with the post-earthquake resistivity response it seems probable, from figure 3, that the resistivity will return to its original value.
In order to quantify the resistivity effects, it was necessary to make these observations continuously for more than 10 years. Morgan and Qian J. (1986) have taken a look at the spatial spreading of the resistivity anomaly over 10 years spanning the earthquake. It is quite interesting to observe the resistivity anomaly grow and then recede. The linear growth rate is of the order of 100 km/yr (^ 0.3 cm/sec). This is a unique data-set in that it shows an earthquake process both temporarily (over 10 years) and spatially (200 km x 100 km).
The strain change accompanying the Tangshan earthquake has been computed to be ^ 10" (Qian J., 1984). Taken with figure 3, showing

_2 resistivity changes of the order 10 , the 10 3 amplification is again
evident. The decrease in resistivity before the earthquake is consistent with compression in the unsaturated surface zone. Water table changes in the Tangshan-Tianjin region were given by Qian J. (1984) and reproduced in figure 12. There is a clear co-seismic signal. It is believed that the regional decreasing trends are not associated

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with the earthquake but rather with excessive water depletion for farming purposes. The slow decreasing water table would cause a resistivity increase rather than the decrease observed in Tangshan.
The Tangshan resistivity anomaly is also well related to the seismic data. Figure 13 from Qian F. (1983), shows fault plane solutions plotted using seismic (dashed line) and electrical (solid line) data. This is a unique display of the data which clearly shows a relationship between the seismic and electrical stress fields. Figure 14 shows the correlation between the aftershock seismicity and the resistivity anomaly (Qian J., 1979).
The data of figure 3 does not show appreciable co-seismic resistivity response. The temporal strain response is not well known for this earthquake, however, an earthquake related strain release of
- 5
~ 10 must have occurred and was indeed responsible for the steps in the water table levels of figure 12. Figure 15, a compilation of other Chinese data is taken from Qian F. et al. (1983). Note the extremely good correspondence between the horizontal strain measurements and the average resistivity response. However, there is again little co-seismic step response, except perhaps for the Songpan earthquake.
The slow variations in the resistivity responses of figure 3 clearly indicate a rate process (Morgan and Qian J., 1986). Rutter and

- 9 -1 Mainprice (1978) showed that for strain rates less than 10 sec , the
"elastic" response is controlled by pressure solution. For the Tangshan

- 13 -*5 earthquake the strain rate must have been ^ 10" /sec (10" strain over
2 to 3 years). Little is known about the effects of pressure solution on resistivity changes. However, pressure solution can indeed alter the pore space porosity both in terms of its magnitude and its microscopic connectivity. Archie's exponent "m" is related to connectivity.
Qian J. (1984) considered a wide variety of possible mechanisms to explain the Tangshan earthquake anomaly. He was unable to foster a reasonable physical explanation. If the ~ 10 3 amplification factor between resistivity and strain is assumed then it is indeed possible to explain not only the resistivity changes of the order of 1%, but also the decrease in resistivity. Recall that this high sensitivity is related to the stress-strain response of cracks or narrow pores.

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Consequently, if a pressure solution mechanism is imposed, it must be taking place in such a manner that it preferentially affects the (narrow) crack porosity. This is not unreasonable since the highest stress levels are likely to be in the narrow cracks.
Unfortunately, the real situation is not this simple. Figure 1, from Bufe (1973a) clearly shows co-strain resistivity steps. Further-more, the Japanese variometer of Yamazaki (1974) has given many examples of co-seismic resistivity changes (figure 16). The former experiment of Bufe, because of its geometry, measured a large component of vertical resistivity. The Japanese variometer is essentially a point measurement because of its one meter electrode spacing. The Tangshan and related resistivity and strain measurements sample mainly the horizontal stress field. Both the borehole strain meters (see for example Galdwin, 1986) and water level changes show co-seismic responses. The resistivity data in figure 3, for Tangshan, are monthly averages as compared to virtually real time measurements for the other cases which show co-seismic responses. However, even at this low sample rate a significant permanent strain change would be discernable. Reconciliation is not clear at the present time, however, it may be that some of our present strain measurements are more sensitive to vertical than to horizontal strains or that more than one physical process is operating. We also need a better understanding of the relaxation time in reservoirs (Bredhoeft et al., 1986). Perhaps resistivity changes not only in response to strain but also in response to stress via the pressure solution mechanism. Precursory resistivity signatures may therefore be accompanied by negligible strain variations, thereby furnishing a unique and distinct precursor. A much larger and longer data-set is needed in order for us to resolve some of these issues.
Summary
There are still a number of aspects of the proposed model that must
be resolved. The resistivity changes must be better correlated with high quality data of strain, measured by a number of different independent means. Moreover, it is important for us to do laboratory experiments at very low strain levels and to observe whether such relaxation phenomena are indeed related to pressure solution effects with coupled resistivity changes.

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TOWARD A METHOD OF PREDICTION

In moving towards a method of prediction using electrical methods a large data-base must be established. There are a wide variety of methods that may be used in monitoring the conductivity. I have tried to emphasize in this article the need for near-surface measurements. The near-surface constraint immediately removes low-frequency magneto-tellurics, tellurics and also long baseline dipole-dipole arrays. High frequency electromagnetic methods will be sensitive to the near-surface structure, however, the required stability and accuracy would be difficult to achieve. The resistivity method is a relatively simple and robust experiment via which high signal-to-noise, stability and accuracy can be attained. Sensitivites to 0.01% or better can be obtained. Recall that Fitterman and Madden (1977) obtained sensitivities of 0.005% over short periods with relatively basic instrumentation and without any further digital signal processing. The details of a suitable instrument system are described in Morgan and Qian J. (1986).
Another critical component is the siting of the resistivity arrays both in terms of the regional geology and the configuration. High sensitivites are to be expected in rocks of low porosity, with a large component of crack porosity. Generally, such materials would have a high resistivity and local resistivity surveying is necessary before implantation of an array. Strong arguments have been made for the near-surface measurements, however, the surface zone also represents the region most susceptible to noise. Noise in this respect comes from temperature, rainfall and cultural electromagnetic sources. There are only practical real solutions to this problem,
(1) do surface measurements with multi-electrode arrays. Invert to remove or deconvolve the surface effects out of the data to expose
the resistivity variation at a depth of say 100 m to 500 m.
(2) Conduct measurements in a borehole or boreholes significantly deeper than the water table depth.
There are pros and cons to either method, however, until they are tried extensively we will not be able to weigh their relative merits.

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Summary
Based on past experience in China, Japan and the USA, and coupled with our preliminary model, it is recommended that the best electrical measurement for earthquake prediction purposes is an active resistivity measurement sensing the near-surface zone.

Conclusions
The future looks good for the use of geoelectricity in earthquake prediction. A reasonable theory is emerging based on our present knowledge of rock physics. Moreover, what we learn from the electrical behavior may influence the broader ideas of the mechanical processes producing earthquakes. Much more data is needed, both from the laboratory and the field. It is strongly suggested that appropriate resistivity measurements be established on the San Andreas, perhaps Parkfield, as soon as possible.

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LINEAR STRAIN
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Rock
Expansion

Increasing
Resistivity

NOAA STONE CANYON GEOPHYSICAL OBSERVATORY FEB 27 1972

Figure 1: Strain and resistivity variations as recorded by Bufe (1973a).
(a) Resistivity and strain at Stone Canyon, six kilometers from a magnitude 5 earthquake.
(b) Strain and resistivity steps at STC from local earthquakes.

Figure 3: The curves of monthly mean of resistivity in the Tangshan area recorded continuously at 6 stations around the epicenter
since their establishment (from Qian F., 1983).

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Figure 4: Contoured observed resistivity variations preceding the Tangshan earthquake of China. Data from Qian J. (1985).

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DISPLACEMENT
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SHEAR STRESS
ON
SURFACE

Figure 8: Discrete screw-dislocation model of a right lateral strike-slip fault.
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z (km)

Figure 10: Approximate extent of regions showing substantial resistivity change. The fault is assumed to be locked to 3 km.

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Ap/po (1 div. = 10%]

7 (1 div. = 2.5 bars)

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Figure 11: Changes of electrical resistivity, pore pressure, and shear stress for Franciscan sandstone sliding on Gabilan granite.
Normal stress was kept at 500 bars (from Wang et al., 1978).

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Shuangqiao

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c)

b)

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No.2»
No.1

°Shuangpiao

°Baitangkou

OACHENG

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Figure 12: Watertable changes in Tangshan-Tianjin region: (a) data from wells in Tangshan region; (b) data from wells in Tianjin

region; (c) the locations of wells in (a) and (b) (from Qian

J., 1984).

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Figure 13; Relation between the earthquake source mechanism and distribu-tion of sign of p a before Tangshan earthquake; in which aa, a
bb; a'a', b'b' are possibly the nodal lines; shaded sectors
represent regions where the first P-arrivals are "-"; geoelectric station numbers are the same as those in Figure
2. The locations of geoelectric stations represent the
projections of the stations on the upper hemispherical surface.

(a) Stations with drop of p a

a

(b) Stations with drop of p a and interruptions in data a (c) Stations with rise or no change in p a
(from Qian F. et al., 1983)

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Figure 14:
(a) The distribution of earthquakes (M > 4) during the period after the great Tangshan earthquake (July 28 - Nov.
30, 1976). Open circles indicate aftershocks, solid circles indicate the earthquakes that occurred in the region beyond
the aftershock area.
(b) The contour map of anomalous strength of P Q in July, 1976, in the Beijing-Tianjin-Tangshan region. The numerals on the contour lines denote the pe
Adapted from Qian J., 1984.

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May 16. 1968

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Figure 16: Resistivity step for earthquake No. 1 as recorded by hfgh, H, and low, L, sensitivity channels. The observation point, A,
and epicenter are indicated on the map with a small circle
and a cross respectively (from Yamazaki, 1974).

Tectonomagnetism and Intermediate-Term Stress Monitoring

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Two extensive arrays of proton precession magnetometers deployed in different tectonic set-tings have been investigated for the significance of piezomagnetic stress monitoring in earthquake studies. The first of these has operated since 1973 in California, involving up to 34 continuously monitored sites. The second operated in a region of very high seismicity in the southern Pacific ocean since 1980 and comprised 25 sites. The arrays have provided considerable insight into the capabilities and limitations of piezomagnetic monitoring which on well documented physical prin-ciples and extensive laboratory studies was expected to provide a reasonable basis for large scale intermediate term stress monitoring.

It is now clear that these experiments have provided continuous high resolution data at the sensitivities originally proposed, that co-seismic signals have been observed and that regional strain anomalies of tectonic causes can be effectively mapped using piezomagnetic anomalies. However, few examples of intermediate or short term precursors to earthquakes have been obtained. The primary reasons for this appear to derive from accumulating evidence that the ambient shear stress near active faults and the mean deviatoric stress release prior to and during earthquakes is lower than than had been assumed when the studies were initiated and that the observational limits of magnetometers separated by about 10 km, after application of a variety of noise reduction techniques, appear to be no better than about 0.2 nT. Lower estimates of the average change in deviatoric stress associated with earthquake rupture correspondingly reduce the fields calculated from tectonomagnetic models.

Some unexpected and exciting results have been obtained which indicate many potentially useful aspects of magnetic field changes near active faults are still not completely understood. The more important of these results are: 1) the observation of correlated magnetic, strain, uplift, and gravity changes during the latter stages of a remarkable deformation episode near the San Andreas fault in southern California, 2) indications of anomalous change in magnetic fields along the San Andreas fault to the southeast of Palm Springs, 3) the first observations of coseismic seismomagnetic effects on two independent magnetometers during the July 8, 1986, ML 5.9 North Palm Springs earthquake.

BACKGROUND

Tectonomagnetism - the study of magnetic fields that result from earthquake or volcano related crustal deformation - has long shown promise as a simple and inexpensive method for monitoring deviatoric crustal stress and perhaps providing a tool for predicting crustal failure (Wilson, 1922; Kalashnikov, 1954; Stacey, 1964; Stacey et al., 1965), but few unambiguous results have been obtained. Subfields of tectonomagnetism are volcanomagnetism and seismomagnetism which describe the magnetic field perti rbations related to the actual volcanic eruption process and the earthquake rupture process, respectively.

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The stress dependence of rock magnetization has been demonstrated under laboratory condi-tions (Kalashnikov and Kapitsa, 1952; Katipsa, 1955; Ohnaka and Kinoshita, 1968; Kean et ai, 1976; Revol et ai, 1977; Martin, 1980; Pike et ai, 1981) and theoretical models have been developed in terms of single domain and pseudo-single domain rotation (Kern, 1961; Stacey, 1962; Nagata, 1969; Stacey and Johnston, 1972) and multi-domain wall translation (Kern, 1961; Kean et ai, 1976; Revol et ai, 1977).

The stress sensitivity of induced and remanent magnetization from theoretical.and experi-mental studies have been combined with stress estimates from dislocation theory of fault rupture and elastic pressure loading in active volcanoes to calculate field changes expected to accompany earthquakes and volcanoes (Stacey, 1964; Stacey et ai, 1965; Johnston, 1978; Davis et ai, 1979; Sasai, 1980; Hao et ai, 1982; Davis et ai, 1984). These models show that moderate scale magnetic anomalies of a few nanoTeslas (nT) should be expected to accompany earthquakes and volcanic eruptions for rock magnetizations and stress sensitivities of 1 Ampere/meter (A/m) and 10"4 bar" 1, respectively. Unambiguous observation of local magnetic signals of this amplitude require dedi-cated efforts to understand and reduce noise in magnetic field measurements. An intrinsic appeal of the technique is that piezomagnetic anomalies are stress induced, and hence in principle allow monitoring networks to isolate stress accumulations from simple strain changes accompanying creep processes in the earth.

TECTONOMAGNETISM

a) Crustal Deformation

A very notable result for intermediate term monitoring studies obtained during the past four years has been the identification of temporal magnetic fields that correspond to local gravity, strain, and uplift changes during part of a remarkable episode of crustal deformation in southern California. This deformational episode, known generally as the "Palmdale Uplift", was first identified in leveling data (Castle et ai, 1976). During the period between 1974 and 1976 changes in the uplift in the region between Cajon Pass and Palmdale occurred at the same time as offsets in local magnetic field (Johnston et ai, 1979). Aseismic strain changes were also identified in geo-detic data taken near the San Andreas fault between 1979 and 1982 (Savage et ai, 1981a, 1981b; Savage and Gu, 1985) and, in an attempt to intergrate these and other data, Jachens et ai, (1983) reported that, during the 1979 to 1982 period, changes in level lines, changes in strain, and changes in gravity, all occurred in a correlated manner.

When continuous measurements of magnetic field in each of the regions investigated by Jachens were sampled at the same time as the gravity strain, and level data, and then tested for correlation, the results showed correlation significant at the 95% level or better (Johnston, 1986). Figure 1 shows the regression plot between gravity data and magnetic data from the Tejon, Palm-dale and Cajon regions. When the data are plotted all together in this way (i.e. with no allowance for different response in different regions) the correlation coefficient is 0.93 and is significant at the 1% level. Figure 2 is a composite plot showing superimposed time-histories of magnetic field (stars), gravity (dots), areal strain (triangles), and elevation (squares) data, together with their error bars, from the three regions in southern California.

Least-square fits between the magnetic data and each of the various parameters give transfer functions of the form;

AF=-0.03A0 (1)

where AF is in nanoTeslas and A0 is in microGals,

AF=-0.98A«fratn (2)

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+1.0-

-20

Ag (/U gals)

Fig. 1. Plot of magnetic field against gravity fluctuations from the Tejon, Palmdale, and
Cajon regions in southern California. Shown also is the least-squares fit to the data (from Johns-ton ct a/., 1986).

where Astrain is areal strain in ppm (compression negative), and

AF =9.1 Aft (3)

wh«re Aft is in meters.

Two possible explanations for these relationships exist. Either a short term deformational episode has been independently recorded in each data set or all data sets have been contaminated by a common source of meteorologically generated crustal or instrumental noise. Because the inferred relationships between these parameters is in approximate agreement with those expected from simple deformational models (Jachcns ct ai, 1983) and tectonomagnetic models (Johnston, 1986), and since no relation was found between the continuous magnetic field data and either rainfall, pressure, or temperature, the deformational explanation is preferred (Johnston, 1986). Correspondence between leveling data and magnetic field changes have been also observed in Japan associated with an episode of uplift on the Izu Peninsula (Ohshiman ct ai, 1983; Honkura and Taira, 1983).

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b) The Hollister 1974 Earthquake

For the two data arrays, the only possible precursor identified related to the Hollister, November 28, magnitude 5.2 earthquake of 1974. A systematic increase in magnetic field of 0.9 nTesla occurred at a magnetometer site 11 km from the epicentre during the early part of 1974. A more dramatic increase on 1.5 nTeslas occurred about seven weeks before the earthquake and lasted for about two weeks. Four weeks prior to the earthquake, the magnetic field returned to approximately its initial value, and remained at this value through to April 1975. These data cannot be explained by ionospheric disturbances or telluric currents. The anomaly was demon-strated to be present at a much reduced level at another site 15 km from the epicentre (Davis et a/., 1980). The most probable cause for the anomaly is the piezomagnetic effect, but since it is not possible to easily relate the time history of the anomaly to known fault motions in this time interval, the anomaly's only value is as a precursor to this particular event. The stress sensitivity of the technique is supported by the event, but implications regarding stress redistribution with time and fault interactions are poorly understood.

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Fig. 2. Superimposed time-histories of magnetic field (stars), gravity (dots), areal strain (tri-
angles), and elevation (squares) data, together with their error bars, from Tejon, Pajmdale, and
Cajon. Shown also is the least-square linear fit to the data (from Johntton, 1986). 284

c) South Pacific Array
Two major earthquakes occurred within 50 km of at least two monitoring sites. Neither event produced co-seismic offsets at any station at the 0.5 nTesla level, though anomalies could reasonably have been expected from both events. In the twelve months prior to the two earth-quakes, however, all stations within 200 km of the epicentres had experienced anomalies which were well above the noise, and could be interpreted as a slow propagation stress episode moving north to south across the array. The offsets did not coincide with significant ionospheric distur-bances, and all sites showed maximum calibration errors of less than 0.5 nTesla at subsequent maintenance visits. These offsets are probably of piezomagnetic origin, but again bear no simple relationship to present understandings of fault dynamics and interactions.

d) Instrument Development and Measurement Precision

Determination of the precision of local magnetic field measurements with arrays of magne-tometers as a function of instrument, spatial scale, sampling frequency, and site location has attracted continued attention since 1982. Simple differences from 5 km to 10 km station separa-tions are quite effective at reducing disturbance fields from external geomagnetic field variations common to all sites. However, the ability to resolve temporal magnetic fields of crustal origin becomes progressively worse with increasing site separation, increasing time duration, and with differences in magnetization between sites. The decreased resolution with increasing site separa-tion measured on the 27 recording magnetometers installed along 800 km of the San Andreas fault has the form;

ff(nT)=0.01(±0.003)D (Arm)+0.07(±0.08) (4)

over the spatial scale of most interest for tectonic processes of 0 to 53 km. a is the standard devi-ation of hour averages of magnetic field differences (Johnston et a/., 1984). This linear increase in

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FREQUENCY (cpd)

Fig. 3. Geomagnetic difference-field noise spectra obtained from multiple magnetometer
pairs with site separations between 10 and 15 km. Shown also is the observed instrument precision limit obtained from magnetometers with a o.25 nT sensitivity (from Johnston et a/., 1984). Piezo-magnetic monitoring is not limited by instrument precision.

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standard deviation with station separation was not observed in the south Pacific array, where the noise limitation expressed in equivalent form was

<7(n!T)=0.0004(+-0.0007)I> (Jbm)+0.34(±0.12). (5)

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Fig. 4. Comparative power spectral density of simultaneous differences from proton magne-tometers with a 0.12 nT sensitivity and self-calibrating rubidium magnetometers with a 0.014 nT sensitivity. Both sets of magnetometers were installed in a tectonically active region near the San Andreas fault with a site separation of 13 km (from Ware et ai, 1985).
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Power spectra obtained in California from pairs of typical magnetometers with ;site separa-
tions between 8 and 15 km show noise power decreases with increasing frequency by about 8 dB per decade of frequency, as shown in Figure 3. It is only at high frequencies (more than 10 cpd in this case) that the noise power approaches the least count noise limit of the magnetometers (0.25 nT). At lower frequencies these spectra have dominant noise peaks that result from diurnal varia-tions and tidal effects from ocean tide induction (Johnston et ai, 1983).

A comparison of proton and self-calibrating rubidium magnetometers with 0.01 nT sensi-tivity and proton magnetometers with 0.12 nT sensitivity over a range of baselines in seismically inactive and seismically active regions has been made to determine whether and under what con-ditions improved sensitivity might be useful in tectonomagnetic measurements (Ware et ai, 1985). The most relevant experiment using a 13 km baseline in a seismically active and geologically com-plex region of the San Andreas fault showed that external noise which decreases at about 10 dB per decade of frequency, dominates the observations at periods longer than about 10 minutes and both systems make equivalent measurements. This is shown in the comparative plots of power spectral density in Figure 4. For periods less than 3.5 minutes (420 cycles/day), the proton mag-netometers become limited by least-count noise (0.12 nT) and the improved sensitivity of the rubidium magnetometer becomes apparent. If noise power continues to decrease with frequency magnetic field measurements at about the 0.001 nT level should be possible at frequencies greater than 1 Hz. This raises the intriguing possibility that magnetic variation corresponding to the pro-pagation of dynamic seismic waves might be detectable on ultrasensitive magnetometers.
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Fig. 5. Residual total secular magnetic field changes along the San Andreas fault in central
and southern California not apparently related to sources in the core or the ionosphere/magnetosphere (from Johns ton tt ai, 1985). This figure (prepared December,1984) shows a clear anomaly in the vicinity of the Palm Springs July 8 Palm Springs Earthquake, and regionally identifies the Coyote Lake, Morgan Hill, Quiensabe and Mt. Lewis earthquakes.

With regard to the detection threshold for tectonomagentic effects, the most important development concerned the realization that most of the noise (long period and short period) can be removed if corrections are made for local site response to external field variations. Since different sites will respond differently because disturbance fields are in different directions and since disturbance fields in the same direction may generate different responses at different sites because the site magnetization vectors are different, these response effects limit the detection threshold. These effects can be identified and removed by defining multichannel Wiener filters which predict the total magnetic field at a site from component magnetic data during periods of disturbed geomagnetic field (Davit tt al, 1979; Davit et al, 1980; Davit et ai, 1981; Davit and Johntton, 1983). This technique reduced the noise in hour averages from 0.7 nT to 0.3 nT for baselines from 8 km to 100 km (Davit and Johntton, 1983).
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c) Short Wavelength Secular Variation

Further application of this technique was made to correct for site response and identify the secular variation in data from the array of 34 total field magnetometers throughout central and southern California during the period 1976 to 1984 (Johnston tt ai. 1985). The secular variation obtained has the form;

F^k^O+kt't+K (6)

where F is in nT/a, 0 and ^ are the geographic latitude and longitude, and *,, k 2 and K are 1.50±0.08 nT/a.degree, -0.23±0.06 nT/a.degree, and -129.2±0.1 nT/a, respectively.

Surprisingly, the observed secular variation obtained did not agree with that predicted from global secular variation models. The secular variation was removed from the data to attempt to identify crustal fields of tectonic origin. The residual fields obtained are shown in Figure 5 first prepared in September, 1984, and are most apparent on the San Andreas fault in southern Cali-fornia where subsequently a MX, 5.9 occurred (see below). This regional stress anomaly mapping provides a new dimension in intermediate term prediction studies by identification of areas of stress accumulation. The technique is insensitive to strain, and could be performed on a routine basis to identify changes to the large scale stress regime in a regional sense. The areas identified in figure 5 account for .most of the major events in the San Andreas system since 1984.

SEISMOMAGNETISM

a) North Palm Springs Earthquake

A most important result in the field of seismomagnetism was the first recording of co-seismic seismomagnetic effects during the July 8, 1986, North Palm Springs earthquake. This earthquake had a moment of 2*1028 dyne-cm and a magnitude of 5.9. Two total field proton mag-netometers were installed at distances of 3 km and 8 km from the subsequent earthquake and had been sampling once every 10 minutes since early 1979. The data are transmitted with digital telemetry to Menlo Park, California (Muctttr tt al, 1981).

The local magnetic field at the magnetometer closest to the earthquake decreased by 1.2 nT while that at the second, eight kilometers from the epicenter, increased by 0.33 nT. Both instru-ments were on the same side of the fault. Figure 6 shows the records from the two sites OCHM and LSBM referenced to another site in the area CHUM during the period June 1 to July 12, 1986, and also the difference between the two local sites OCHM and LSBM.

A tectonomagnetic model of the earthquake has been constructed using seismic ally deter-mined parameters for the rupture length, width, and the depth. To satisfy the moment a slip of 20 cm was assumed and the magnetization in the region was estimated from surface samples and regional magnetic anomalies to be 2 A/m. Within the uncertainties this model predicts the changes observed (Johnston and Mvtlltr, 1986). The observation confirms the stress sensitivity of the tectoniomagnetic effect in large scale phenomena, at the sensitivity expected from theory and laboratory values.

TECTONOMAGNETIC MODELS

No new work has been reported on tectonomagnetic models but results from related fields may have a serious impact on interpretation of these calculations. New work of the state of stress in seismic ally active areas (McGarr tt ai, 1982) and observations of measured stress drops for earthquakes (Hanks, 1980) suggests the average deviatoric stress change during earthquakes may be about 10 or 20 bars rather than 100 bars as had been previously assumed (Staety, 1964; Sasai, 1980; Hao tt ai, 1982). If stress changes are at this level, then the fields expected from them as a consequence of the piezomagnetic properties of rocks are correspondingly reduced. This, plus the fact that the present tectonomagnetic models already underestimate observations of tectonomag-netic effects generated by dam loading (Davit and Statty, 1972) and uplift (Ohshiman tt ai, 1983; Johnston, 1986) by at least a factor of five, indicates possible inadequacies in the current theory and modeling techniques.

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Fig. 6. Magnetic field at sites LSBM and OCHM with respect to a common site CHUM (OCCH - bottom, and CHLS - middle) as a function of time from June 1 to July, 12, 1986, and also the difference between the two local sites (OCLS, top). The ML 5.9 North Palm Springs earthquake (occurrence time shown with arrow) occurred on July 8, 1986, at an epicentral dis-tance of 3 km from OCHM and 8 km from LSBM (from Johnston and Mucllcr, 1986).

CONCLUSIONS

Extensive array studies using proton precession magnetometers have shown that though the physical basis of piezomagnetic stress monitoring has been verified in field experiments, the value of the technique to intermediate earthquake prediction studies is limited for small stress changes by extraneous electromagnetic noise sources. Recent studies on estimates of the stress drop in earthquakes have lowered reasonable expectations on the stress accumulation which may be

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expected in months to weeks before an earthquake. Piezomagnetic monitoring remains the only stress monitoring technique which has been successfully implemented and has demonstrated con-siderable potential in identification of regional stress anomalies associated with earthquake genesis. Our decision to close operation of the arrays is based on the limited resources available to the subject and on an emerging appreciation of the efficacy of continuous borehole strain array studies for improved insight into the earthquake process. Piezomagnetic monitoring in volcanic studies where it has demonstrated success in distinguishing strain cycles which are accompanied by stress accumulation from those which are not should be continued.

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Temporal Change in Scattering and Attenuation Associated with the Earthquake Occurrence
-A Review of Recent Studies on Coda Waves

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Abstract
The study of coda waves has recently attracted increasing attention from seismologists. This is due to the fact that it is viewed as a new means by which the stress accumulation stage preceding a large earthquake can be measured, since the scattering paths nearly uniformly coyer a fairly large region around the focus and observation stations compared with the direct ray paths. So far, we have had many reports on the temporal variation of the relation between coda duration and amplitude magnitude, and that of the coda attenuation Qc" 1 - Some of these have . shown a precursor-like behavior; however, others seem to have shown a coseismic change. We have critically reviewed these reports, and discussed what these observational facts tell us about the change in the heterogeneous crust. We found significant temporal variations not only in the mean but also in the scatter of Qc" 1 associated with the mainshock occurrence. The formation of new cracks, the reopening and growing of existing cracks, the interaction of these cracks, and the pore water movement through these cracks might correspond to such variations. In addition, we may expect an inhomogeneous distribution of crack clusters in a fairly large region compared with the aftershock region. The gradual appearance of such crack clusters seems to be the most plausible mechanism by which coda decay gradients are caused to largely scatter in the stress accumulation stage.

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/. Introduction
The decay rate of coda amplitude is in general very stable, and such a trait is common to all local earthquakes located near a given station [Aki, 1969]. Coda waves have been interpreted as scattered S wavelets due to heterogeneities in the lithoshpere. Measurements of scattering coefficient (g km' 1) [Sato, 1978] and coda attenuation (Qc' 1) [Rautian and Khalturin, 1978; Tsujiura, 1978], which characterize coda excitation strength and coda decay gradient, respectively, have been made based on the single scattering assumption [Aki and Chouet, 1975; Sato, 1977; Kopnichev, 1975] or the diffusion model [Wesley, 1965]. Assuming single isotropic scattering, and uniform distribution of scatterers and S wave attenuation Qs' 1, Sato [1977] predicted the space-time distribution of coda energy density as
E sis(f:r 0,t)=[W0 g o/(47T r 0 2)]K (0 O t/r 0)exp(-Qc - 1 2 n f t) for 0 O t>r 0, where K (x)- (1/x) In [(x+1)/(x- 1)] ;
f, frequency; r 0, hypocentral distance; t, lapse time measured from the origin time; W 0, S wave source energy; go, total scattering coefficient; y3 0, average S wave velocity (see Figure 1). The asymptote of this function is
E SIS (f:ro,t) - [W 0 g o/(27r £ 0 2 t 2)]exp(-Qc - a 27T ft) for y3 0 t»r0. This coincides with Aki and Chouet's single back scattering model [1975]. The equality Qc' 1 ^ Qs' 1 holds good in this approximation and has been accepted in many reports; however, let us treat Qc' 1 as a phenomenological parameter characterizing coda decay in the following. These functions have been commonly used as the geometrical reference functions for the estimation of Qc' 1. Regional differences in Qc" 1 have been studied in various areas in the world in relation to tectonic activity [Singh and Herrmann, 1983; Sato, 1986b]. The spectral structure of the inhomogeneity has been inferred from the frequency dependent measurements of the scattering coefficient and the P and S wave attenuation [Aki, 1980; Wu, 1982; Sato, 1984a, b].
Chouet [1979] first paid attention to the temporal change in
Qc' 1 for small earthquakes in California. Recently, the study of coda waves has attracted increasing attention from seismologists, since it is viewed as a new means of measuring the stress accumulation for the purpose of earthquake prediction [Aki, 1985]. This new way of measuring stress accumulation has been thought effective, since the scattering paths do not one-dimensionally but three-dimensionally cover a fairly wide region around the focus and observation stations nearly uniformly. So far, we have had many reports on the temporal variation of scattering and attenuation associated with the earthquake occurrence based on coda wave analysis [Gusev and Lemzikov, 1980, 1984, 1985; Jin, 1981; Tsukuda, 1985; Novelo-Casanova et al. , 1985; Jin and Aki, 1986; Sato, 1986a, c; Sato et al. , 1986; Lee et al. , 1986; Peng et al., 1986; etc.]. Some of them showed a precursor-like behavior; however, others seem to have shown a coseismic change. Reported studies were principally based on two kinds of analyses: the comparison between coda

duration and amplitude magnitude, and the spectral measurements of Qc" 1 - Here, we have critically reviewed these reports, and discussed what
these observations tell us about the change in the earth medium.

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2. Recent studies on coda waves
2-1. Relation between coda duration and single station magnitude Phenomenological studies of the empirical relation between coda duration and amplitude magnitude have been intensively done.
A temporal change in the relation was reported for the Tangshan earthquake (M s =7.8. Jul. 28, 1976) by Jin [1981], and Jin and Aki [1986]. Coda durations for earthquakes of the same magnitude were abnormally shorter in a three-year period preceding the mainshock compared with those before and after that period. On the other hand, from studying the Jianchuan earthquake (M L = 5.3, Jul. 3, 1982) in Yunnan, Van and Mo [1982] found that coda durations were longer for the foreshocks than those for the aftershocks for the same maximum amplitude of S waves. Gao [1985] briefly reviewed recent studies of this kind made in China.
Hasegawa and Hori [1985] analyzed vertical component coda at a station just above the focal area of the southeastern Akita earthquake (Ms-6.2, Oct. 16, 1970) in Japan. The mainshock focus was 6km in depth. They found that the maximum amplitudes for the P and S waves of the foreshocks were greater than those for the aftershocks of the same coda duration for 10- 15Hz. By assuming a constant scattering coefficient, they concluded that the attenuation above the mainshock focus increased after the mainshock occurrence.
In these studies, magnitudes were calculated from the maximum amplitudes of direct waves at a well calibrated single station. Therefore, there remains a possibility that the reported change might be caused by the change in focal mechanisms, not by the real change of attenuation in the earth medium.

2-2. Relation between coda duration and average magnitude
In order to avoid the affection of radiation pattern changes, it is necessary to investigate the relation between coda duration and average magnitude, which is the arithmetic average of station magnitudes at the stations surrounding the epicenter.
- The eastern Yamanashi earthquake (Ms =6.0, Aug. 8. 1983) in Japan-Sato [1986a] found that coda durations (f-o time) at two stations,
TRU and HHR, in the close vicinity of the aftershock region were shorter before than those after the mainshock occurrence for the same average magnitude. For the analysis, he used earthquakes which took place in a small volume of dimensions 10x10x5 km 3 (hatched region in Figure 2) near the mainshock hypocenter. Station magnitudes were calculated from the maximum amplitudes of vertical seisrnograms irrespective of the P or S waves. Especially, TRU is located just above the aftershock region and equipped with a seismometer at the bottom of a well of 160m in depth. Such differences were significant for coda durations shorter than about 30s (see Figure 3), which corresponds to a travel radius of about 50km. The residuals of station magnitude from average magnitude were lower before the mainshock than after at five stations, ENZ, TRU, HHR, AKW and ASG, in the vicinity of the aftershock area; however, changes in the residuals of station magnitude at two farther stations, SMB and JIZ, took the opposite sign. It is difficult to explain these changes by the change in focal mechanisms.

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From the two independent analyses, Sato concluded that the S wave attenuation intensity was higher before the mainshock than after within a 26km radius above the mainshock focus, which was 18km in depth, in the upper crust as illustrated by the speckled hemisphere in Figure 2: #Qs~ 1 = -1. 4xlO~ 2 f~ l for f-2~20Hz. The predominant frequencies were 2~3Hz around the end of coda and 10~20Hz for direct waves.
There were other reports of precursor-like changes associated with
this earthquake. Figure 4 shows the temporal variation of the ratio Pi for the number of earthquakes of 2<M L <2.5 to that of 1.5<M L <2 around the mainshock focus. The Pi value was higher, that is, the b value was lower, during about a one year period before the mainshock than that before and after this period [Imoto and Ishiguro, 1986]. Qc" 1 was negatively correlated with the b value. Tilt anomaly was observed 18 days preceding the mainshock at ENZ, which is located in an active fault at a 31km distance from the mainshock epicenter [Sato et al. , 1984].
- The western Nagano earthquake (Ms =6. 8, Sep. 14, 1984) in Japan-
Sato [1986c] studied the relation between coda duration at GER close to the mainshock epicenter (A=24km) and average magnitude (see Figure 5). The vertical seismograms of earthquakes which took place in a small restricted volume of 6x6x5km 3 (surrounded by a broken line in Figure 5 (a)) in the middle of the aftershock region were analyzed for a period from Feb. 1982 to Dec. 1984. Coda durations for earthquakes of the same average magnitude were abnormally longer in period B, which began about 16 months before and continued until 7 days after the mainshock, than those in the preceding time period A and in the following period C (see Figure 6). Figure 7 shows the temporal plots of the residual of the logarithm of coda duration from the linear regression. According to the one-sided statistical t-test, the significance level is less than 0.1% where the null hypothesis is that the means of duration residuals for two samples are equal. A gradual decrease is found even in period C. The predominant frequencies were about 5Hz at the end of coda. The temporal plots of the residual of station magnitude at GER from average magnitude scattered largely, however, did not show any systematic variations. It means that there was little systematic change in focal mechanisms or attenuation intensity along the rays between GER and the middle of the aftershock region.
Sato concluded that the most plausible mechanism for the elongation of coda durations was the increase of scattering intensity in the crust within a radius of about 50km from the mainshock focus, since the anomaly was significant for coda durations shorter than about 30s as shown in Figure 6.
Yokoyama [1986] reported that earthquake swarms began to concentrate close to the mainshock epicenter after 1978. Sugisaki and Sugiura [1986] reported an anomalous appearance of H 2 in bubble gases at a spring (A-52km) beginning one month before the mainshock. Katoh et al. [1986] reported that an anomalous increase of radon concentration in soil gas began from 1982 to the summer of 1983 at three sites around the mainshock epicenter, of which the epicentral distances were 25, 100 and 100km, respectively, as illusatrated in Figure 8. The chronological coincidence between the radon anomaly and the coda duration anomaly in the earthquake preparation stage

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suggests that the formation of new cracks and the reopening and growing of existing cracks progressed not only in the close vicinity of the focal area but in a fairly large area. Scattering is strong when the wavelength is comparable to the dimension of a scatterer in general. Even though the length of each crack is much smaller than 1 meter, if we imagine a cluster of cracks with dimension on the order of several hundred meters, the resulting contrast in elastic parameters functions as an effective scatterer for S waves of frequencies around 5Hz. An appearance of such clusters in the stress accumulation stage might elongate coda durations.
Plotting the half period of initial P waves observed at GER against travel time, Ohtake [1986] found that Q P -1 in the focal region was more than twice that of the surrounding region, not only for the post seismic stage of 1.3 years, but also for a 2.5 year period prior to the mainshock, as shown in Figure 9. Ohtake did not find any temporal increase of Qp" 1 in the 16 month period before the mainshock. The highest value appeared for from one day to two months following the mainshock, period A-II, following which the Qp" 1 monotonously decreased approaching the value of the surrounding area. The latter fact coincides well with Shibuya's [1979] investigation for the temporal change of Qp" 1 in the focal area of the off-Izu earthquake (M s = 6.9, May 9, 1974) in Japan for the aftershock stage.

2-3. Coda decay gradient
- The Tangshan earthquake (M s =7.8, Jul. 28, 1976) in China-
Jin and Aki [1986] analyzed vertical seismograms at station
PG, which is at a 120km distance from the mainshock epicenter, based on the single isotropic scattering model [Sato, 1977]. They estimated that Q c " 1= 5xlO" 3 around 2Hz for Feb. 1969- Oct. 1972, Qc" 1= 1.4xlO~ 2 for coda just after the S wave, but 3.2xlO~ 3 for a latter part of coda for Apr. 1973- Feb. 1976, a three year period before the mainshock, and Q c ~ 1 = 4xlO" 3 for Sep. 1976- Jan. 1978 as listed in Table 1. They concluded that Qc" 1 was anomalously high in an elliptic area 100km wide, that is, the aftershock area plus the area extending to the NW direction in the three year period, but that the Qc" 1 of a broad region outside that area became lower than normal. In this abnormal period, the b value took a lower value in a 250x250km 2 area extending to the SW of the aftershock area.
Analyzing the P wave spectra at two stations to the WNW of the mainshock epicenter, Zhu et al. [1977] found that the stress drops of small earthquakes were higher before than those after the mainshock. Supposing frequency independent Qp" 1 for l~10Hz, they also found the increase of Qp" 1: Qp'^lO" 3 before and 1.5xlO" 3 after the mainshock. This brings us to mind the report of a high Qp" 1 around 20Hz in the earthquake swarm region compared with that in the surrounding region for the Matsushiro earthquake swarm in Japan [Suzuki, 1971]. What then does the different sign between the changes of Qp" 1 and Qc' 1 mean?

A big difference in Qc" 1 appeared in coda just after the direct S wave. We are afraid of the affection of focal mechanism changes for such a time window. Precisely examining Table 1 of Jin and Aki [1986], we can test whether the standard deviation of Qc" 1 in the

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three year period was larger than that in the surrounding time periods or not based on the one-sided statistical F-test. The significance levels for the rejection of the null hypothesis are 41.0% and 16.4% for the formar pair and the latter pair, respectively (see Table 2).
- The Haicheng earthquake (Ms =7. 3, Feb. 4, 1975) in China-
Analyzing coda decays at SHE station in the vicinity of the mainshock epicenter, Jin and Aki [1986] estimated Q c ~ 1 ~2xlO~ 3 before and 1.4xlO~ 3 after the mainshock around 7 Hz, as shown in Figure 10. In this case, the difference appeared even in the latter part of coda. The travel radius of the coda was estimated to be about 60km, much larger than the size of the aftershock area. They found a decrease of Qc' 1 at another station, HEN, too. The negative correlation between the b value and Qc' 1 held good also in this case.
Analyzing the P wave spectra, Zhu et al. [1977] reported that
Qp~ 1 ~2xlO~ 3 almost uniformly around the focal region before and after the mainshock, but that Q?' 1 increased to 6xlO~ 3 only in the SW area of the mainshock epicenter, as shown in Figure 11. Analyzing the half period of initial P waves at a station in the close vicinity of the mainshock epicenter, Zhuo et al. [1985] reported an increase of Q P ~! after the mainshock. Here again a discrepancy between Qp' 1 and Qc' 1 appeared.
Examining Table 2 of Jin and Aki [1986], we found that the standard deviation of Qc' 1 was larger before than that after the mainshock. The significance levels for the rejection of the null hypothesis are 5% and 0.7% at SHE and HEN, respectively, based on the one-sided F-test (see Table 2).
- The Ust-Kamchatsk earthquake (Ms = 7.8, Dec. 15, 1971) in Kamchatka-Analyzing three-component short period (T 0 =1.2s) seismograms at three stations surrounding the focal area, Gusev and Lemzikov [1985] found an anomalous change in the gradient of the coda envelope decay. The earthquakes used for this analysis were distributed in a 300x200km 2 area covering the aftershock area, where the minimum coda duration used was 70s. They measured the a value which represents the deviation of logarithmic coda decay gradient: a negative a indicates a steep decay corresponding to a large Qc' 1. They did not refer to any theoretical model but took the empirically averaged coda decay curve as the reference time function. They plotted the running means of the 8 successive data, where each datum was an average of all the components at three stations (see Figure 12 (a)). Compared with the relatively stable background level during 1966-1970, a pronounced anomalous decrease and recovery of a was observed during the one year period preceding the mainshock. They reported such a decrease of a , corresponding to an increase of Qc' 1, had been found for two other large earthquakes in the Kuril-Kamchatka zone, but the recovery was not found (Figures 12 (b) and (c)). They concluded that the decrease in the mean of a was a precursor and that the most probable cause was the increase of Qs' 1. They interpreted that a micro-crack formation would start abruptly about one year preceding the mainshock in a large area. Seismic wave energy was spent on crack growth and compression of gas or fluid filled in cracks.

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They also reported an appearance of a heap in coda waves at lapse time around 100s in the anomalous period. Recently, Gusev [1986] radically proposed a seismic emission mechanism [i.e. Rykunov et al. , 1979] for the explanation of this heap phase in coda.
However, we were unable to find an anomalous decrease in plots
of a for vertical and horizontal components at four stations surrounding the Ust-Kamchatsk earthquake epicenter, as shown in Figure 13 [Figure 4 in Gusev and Lemzikov, 1980] . An increase of the standard deviation of a during the period from 1971 to 1973 looks more anomalous to us (see Table 2). Gusev and Lemzikov [1984] reported that the significance level was less than 5% where the null hypothesis was that the standard deviations of a for 1966- 1970 and 1971 are equal.
- The K=13.3 earthquake (ML =5.2, Feb. 26, 1983)
near Garm in the Peter-the-first range, Soviet Central Asia-Earthquakes of the maximum magnitude 5 or 6 have taken place within a several-year interval in the Peter-the-first range in the Tadjikistan region. EW-component band-pass filtered seismograms [Rautian and Khalturin, 1978] have been in operation since 1978 at station Garm, which is close to the mainshock epicenter (A=20km). Sato et al. [1986] measured logfQc^f) for 7 frequency bands from coda decays for lapse times greater than twice the S wave travel time and smaller than 100s (see Figure 14). The earthquakes that occurred in the aftershock region of dimensions 25xllx20km 3 during 1979-1984 were used for the analysis. At 5Hz band, log(Q c " 1 f) took a significantly lower value in the three-year period preceding the mainshock, where a gradual increase began in early 1981 (see Figure 15). They could not find a clear difference between the means of log(Q c ~ 1 f) before and after the mainshock for the other 6 frequency bands. They, however, preferably pointed out a significant difference between the standard deviations of log(Q c ~ 1 f) before and after the mainshock. Standard devations were significantly larger before the mainshock than after for 0.62, 1.25, 2.5, 5 and 18Hz bands based on the one-sided statistical F-test. The significance level for each frequency band is given in Table 2.
From the instability of coda decay, they interpreted the larger standard deviation of log(Qc ~ 1 f) to be due to the appearance of clusters of cracks in a fairly large region in a stress accumulation stage prior to the mainshock failure. Clear decrease in electric resistivity was observed at four stations near the focal area in the middle of 1982 [A. Sidorin, personal communication, 1985]. It suggests a relation with the ground water movement through cracked media.
- The Round Valley earthquake (Ms -5. 7, Nov. 23, 1984) in California-Since Oct. 1978, four earthquakes of magnitude around 6 and preceding several swarms have occurred in the Mammoth Lakes area. Peng et al. [1986] analyzed vertical seismograms of earthquakes which occurred in a period from Apr. 1984 to Jan. 1985. Applying the single scattering model to coda for lapse times of 20-45s and 30-60s, they reported significant temporal changes in the mean of Qc' 1 in relation to the location of the mid point of hypocenter-station pair. Those changes were as follows: (1) Qc' 1 was smaller before the mainshock than after in regions near the mainshock epicenter mostly

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for frequencies 3~12Hz; (2) Qc" 1 was larger before the mainshock than after in regions farther away from the mainshock epicenter; (3) the high Qc" 1 in the Long Valley Caldera which was comparably higher than that for the outside area disappeared after the mainshock. They introduced the doughnut pattern model [Mogi, 1977] for the interpretation, expecting a low Qc" 1 in the area of quiescence and a high Qc" 1 in the active zone. But, the finding of a high Qc" 1 in the outer region runs counter to the case of the Tangshan earthquake.
We found that the standard deviation of Qc" 1 seems to have systematically changed for wide frequency bands from the calculation based on Tables 1-6 of Peng et al. [1986], as shown in Table 2. The standard deviation of Qc" 1 was larger before the mainshock than after in the Long Valley Caldera, region I. From the study of Qc" 1 in different time windows, the standard deviation of Qc" 1 before the mainshock was larger than those after the mainshock for 30-60s, but the change was opposite for 20-45s, in the regions S and T close to the aftershock region. Coda for the shorter lapse time samples the closer region. The larger standard deviations before the mainshock are seen in Figures 4 and 5 in Lee et al. [1986] as they suggested (see Figure 16). We may say that the coda decay gradients were more unstable before the mainshock than after in a fairly large region outside of the aftershock region; however, they became more unstable after the mainshock than before in the vicinity of the aftershock region.
-Other coda decay studies on natural earthquakes-
Novelo-Casanova et al. [1985] reported high Qc" 1 before and low Qc" 1 after the Petatlan earthquake (M s = 7.6, Mar. 14, 1979) in Mexico at 6Hz band. We are afraid of the strong affection of the radiation pattern changes, since they applied the single isotropic scattering model to the early part of coda for the estimation of Qc" 1.
Tsukuda [1985] analyzed vertical seismograms of micro-earthquakes before and after the Misasa earthquake (M S::I 6.2, Oct. 31, 1983) in Tottori-ken, Japan. He reported that the mean of Qc" 1 at 5Hz band was significantly higher after the mainshock than that during the preceding seven years as shown in Figure 17, where the significance level was less than 22 according to the statistical t-test.
We have summarized the reported changes in the means and the standard deviations of Qc" 1 in Tables 1 and 2.

2-4. Ratio of coda duration of horizontal component
to that of vertical component
Coda durations strongly depend on the magnification scale of seismograms. So, Malamud [1974] began to monitor the nondimensional quantity: the ratio of coda duration of the horizontal component to that of the vertical component. Analyzing earthquakes within 100km in epicentral distance and where the durations were mostly from 100 to 300s, he found that the ratio had a lower value in the few months preceding the major earthquakes of KM3 (M L >5) in Soviet Central Asia, as shown in Figure 18.
Van and Mo [1984] studied the Jianchuan earthquake (M L = 5.3, Jul.
3, 1982) in Yunnan, China. They found a clear decrease of the ratio in the few-day period preceding the mainshock, as plotted in Figure

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19, when coda durations (f-p time) were mostly from 10 to 30s.
These facts suggest the possibility of different values for
Qs' 1 between vertical and horizontal polarizations. Gusev [1986] introduced crack-induced anisotropy of Qs' 1 for the explanation. In addition, the possibility is also suggested that coda amplitudes might be different for different components according to the change in focal mechanism.

2-5. Coda of explosion seismograms
Nikolaev's group at the Institute of the Physics of the Earth, USSR intensively repeated explosions in two lakes, Iskanderkuli and Chashmaisangok, in the western part of the Gissar-Kokshal zone, in Soviet Central Asia from 1978 to 1984. They recorded seismic waves at epicentral distances of 50~130km. Analyzing the power spectrum of the radial component coda waves, Gamburtseva et al. [1983] found an anomalous decrease in the spectrum around 2.5Hz for lapse time of 21-40s prior to the two nearby K~12 (M L ~4.5) earthquakes of 1979 and 1980, as shown in Figure 20. From the same experiments, S. P. Starodubrovskaya [Personal communication, 1985] found the appearance of an anomalous heap in coda waves at lapse time around 40s in a period just before these earthquakes. Such a heap was observed at stations for which the direct ray paths crossed active faults. But, such a heap did not appear in a seismogram at a station for which the direct ray path did not cross the active fault.
In China, Gu et al. [1982] studied the temporal change in coda
of explosion seismograms before and after the Liyang earthquake (M s = 6.0, Jul. 9, 1979) (see Figure 21 (a)). They found anomalies in the relation between the predominant frequency of coda and lapse time, as shown in Figure 21 (b). They found that the exponential decay factor a for the coda envelope was abnormally small (see Figure 21 (c)), that is, there was a small Qc" 1, arid that the coda durations were anomalously long in the one year period prior to the mainshock.

3. Discussion

3-1. Overview of the reported changes in coda waves
We can summarize the above reports as follows:
1. Changes in the relation between coda duration and amplitude magnitude were found to be associated with the earthquake occurrence, where predominant frequencies were relatively low such as 2~6Hz.
2. In most cases, the mean value of Qc' 1 was higher before the mainshock than after, or higher in the few-year period around the mainshock than the surrounding time periods for frequencies mostly lower than or equal to 12Hz, and this change has been interpreted based on the change in Qs" 1. In some cases, however, the change was opposite. Many reports showed a negative correlation between the b value and Q c ~ l.
3. Spatial difference of the change in the mean of Qc' 1 was found to be based on the coda decay analysis for different time windows. The doughnut pattern model was introduced for the explanation.
4. Temporal change in the scatter of Qc" 1 in relation to the earthquake occurrence was systematic for wide frequencies. In most cases, the standard deviation was larger before the mainshock

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than after. The increase in the scatter of Qc" 1 began a few years before and/or continued until a little after the mainshock in some cases. The study of the Round Valley earthquake suggests that the standard deviations of Qc" 1 were larger before than after the mainshock, especially in the outer region of the aftershock region.
5. Abnormal increases in radon concentration and an appearance of H 2 gas were observed at large epicentral distances in a period of long coda durations. A decrease in electric resistivity was observed in the vicinity of the focal region in a period of large scatter of Qc" 1.
6. These observational facts suggest that physical property changed in a fairly large region compared with the aftershock region prior to the mainshock.
However, we are afraid that the above summary might be too optimistic. Consequently we would like to add the following comments:
1. Changes in the relation between coda duration and single station magnitude might be caused by the change in radiation pattern. The change in the mean of Qc" 1 also might be caused by the change in focal mechanism in some cases, since the single isotropic scattering model was applied to the early part of coda close to the direct S wave mostly in the vertical component. Anomaly appeared mostly in small magnitude earthquakes, for which the focal mechanism determination was difficult. Sato [1984a] recently showed that the decay gradient of the coda envelope is different for different components, and strongly depends on the focal mechanism in such a time window based on the single scattering assumption for vector waves. Figures 22 (a),(b) and (c) show synthesized envelopes for a three-component seismogram of a M L =3 earthquake at 30km in distance for three different focal mechanisms, where the randomness of elastic moduli is characterized by the exponential auto-correlation function with 10% in RMS fractional fluctuation and 2km in correlation distance.
2. In some reports, the single scattering model was applied to a latter part of coda of lapse time longer than the mean free time of S waves, which was roughly estimated to be about 30s [Sato, 1978, 1984aj. For such a long lapse time, coda amplitudes are theoretically different in different components for the single scattering assumption, as illustrated in Figure 22. Multiple scattering should be considered in such a lapse time. Gao et al. [1983] constructed a multiple scattering model up to the 7th order. A numerical simulation approach was taken by Frankel and Clay ton [1986] for the case of short mean free path. They demonstrated the discrepancy between Qc" 1 evaluated from coda decay gradients and the scattering attenuation of direct wave amplitude. Gusev and Abukakirov [1986] proposed a reference time function for acoustic energy density interpolating the single scattering model and the diffusion model based on the Monte-Cairo simulation for nonisotropic multiple scattering. Will multiple scattering decrease the difference between the coda amplitudes in different components for long lapse time?
3. Contrary to our expectation, there are very few reports on the temporal change in the mean of Qc" 1 for high frequencies.
4. There are many reports on the increase of QP -I in the close vicinity of the aftershock region after the mainshock occurrence compared with that before. Does the contradiction between the

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temporal changes of Qc" 1 and Q P ~' indicate the physical difference or the spatial difference between the sampled regions ?
5. We should pay more attention to the change in scattering coefficient g, which controls the absolute level of coda excitation. If the scattering coefficient largely increases in a restricted region, coda durations become longer and coda decay garadients of earthquakes in this region become steeper than normal at a nearby station. It may be interpreted as an increase of Qc" 1. In the opposite case, if the scattering coefficient increases larger in the outer region, coda durations become longer but Qc" 1 apparently becomes smaller than normal. Like this, Qc" 1 reflects the spatial inhomogeneity of the scattering coefficient distribution. We should remember that Qc~ 1= Qs -1 holds good only when the spatial distributions of Qs -1 and g are uniform and isotropic, and when the scattering is weak.
6. There is a report of little temporal change in Qc' 1 before and after the Adak earthquake (m b = 5.8, May 6, 1984) in the Aleutian islands [Scherbaum and Kisslinger, 1985]. Poupinet et al. [1984] found earthquakes of very similar wave form before and after the Coyote Lake earthquake (M s =5.9, Aug. 6, 1979) in California.

3-2. Cluster formation hypothesis
We can easily imagine a formation of new cracks, reopening and growing of existing cracks, and interaction and connection of nearby cracks in the stress accumulation stage [Nur, 1972; Scholz et al. , 1973; Mjachkin et al. , 1975]. Crampin et al. [1984] proposed the extensive-dilatancy anisotropy of micro-cracks throughout a fairly wide region around an impending earthquake. It is expected that Q c -1 will detect the change in heterogeneous structure in a broad region including an earthquake source region as suggested from distant precursors observed [Aki, 1985]. There is, however, a contradiction between the both temporal and spatial changes in the mean value of Q c -1 in different reports. Recently, Gusev [1986] proposed different time histories for the dilatancy region around the impending earthquake and the outer region in the compressional stress field to explain the doughnut pattern. So far, we are not sure whether these discrepancies will be settled by the doughnut pattern model or not. Here, let us pay more close attention to the change in the standard deviation of Qc -1, not as a credibility parameter for the mean of Qc" 1 but rather as a physical parameter. Many reports have suggested an abnormal increase in the scatter of Qc -1 for the period just before or just around the mainshock occurrence. In addition, there are reports which have shown the anomalous appearance of a heap in coda. These observational facts suggest the instability of the coda decay gradient in the stress accumulation stage.
Kulagina et al. [1982] showed an anomalous increase in the scatter of the Wadati-diagram plots in the earthquake preparation stage in Soviet Central Asia, as shown in Figure 23. Sobolev [1984] experimentally reported the appearance of high and low velocity spots and clusters of acoustic emission hypocenters, not only in the close vicinity of the fault preparation zone, but also in a wide region of a big granite block under the uniaxial loading before the fracture, as shown in Figure 24 (a). He mentioned that the most prominent precursor

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was the increase of the scatter of seismic velocity as shown in Figure 24 (b). The elastic moduli are inhomogeneous as inferred from well log data [Sato, 1984b], and the in-situ stress distribution is also nonuniform as measured precisely by the hydraulic fracturing experiment in a drilled borehole, as illustrated in Figure 25 [Ikeda and Tsukahara, 1986].
It is very natural for us to imagine a nonuniform cluster formation of cracks in a fairly large region compared with the focal region. The decrease of the b value in this stage is in harmony with the repression of the independent appearance of small cracks. The resulting contrast in elastic parameters functions as a scatterer for S waves. We can expect a wide distribution for the size of clusters. Clusters of the dimension on the order of several hundred meters are able to change the relation between coda duration and magnitude in relatively low frequencies less than lOHz. We can expect the movement of pore water through cracks in such clusters. Then, a cluster of cracks containing water as viscous fluid or as adsorved on the mineral surfaces [Clark et al., 1980] works as a strong intrinsic attenuation spot for S waves. We may expect that the increase of elastic scattering will dominate over the increase of intrinsic and scattering attenuation for the coda wave formation in some cases, and that the opposite situation will appear in the other case.
The gradual appearance of the inhomogeneously distributed scatterers and attenuation spots will make the coda decay gradients largely scatter in the stress accumulation stage. We may interpret these phenomena by analogy with the increase of the fluctuation in the critical state in physics. Just or a little after the rnainshock occurrence, we may expect that the growth of clusters of cracks will stop in the outer region, and then coda decay gradients will become stable. On the other hand, coda decay gradients will be more unstable after the mainshock in the aftershock region because of the mainshock and aftershock fractures and the pore water movement through them.

3-3. Suggestions for further researches
The following coda researches are recommended for a better understanding of the temporal change in inhomogeneities in the crust in relation to earthquake occurrences.
1. Monitoring not only the mean but also the scatter of Qc" 1 itself as a physical parameter. The latter is valuable for examining the cluster formation hypothesis.
2. Precise examination of individual heaps in coda waves.
3. Comparison of coda measurements not only with the b value and stress drops but also with geochemical observation (Rn, H 2,), ground water monitoring, and electric resistivity measurement for examining the crack formation and the pore water contribution.
4. Re-examination of the coda portion of explosion seismograms. It is effective for the quantitative evaluation of the scattering coefficient g.
5. Observation of coda by a three-component seismometer in order to clarify the energy partition into different components, when the simultaneous determination of the focal mechanism is necessary.
6. Comparison between Qs" 1 measured from direct S waves, and Qc" 1

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and g measured from coda waves. Qc~! is probably a complicated function of the spatial distribution of Qs -1 and g.
7. Theoretical construction of the inversion scheme for the spatial distribution of Qs -1 and g from coda envelopes, since different lapse time window samples different region in space. It will be useful to examine whether the doughnut pattern model holds good or not. At first, we should try to make such a scheme for coda of short lapse time based on the single scattering approximation.
8 Theoretical construction of the multiple scattering model for the envelopes of three-component seismograms for long lapse time.
9. Precise measurements of in-situ stress in deep boreholes by using the hydraulic fracturing technique to clarify the inhomogeneous scale in the crust.

Acknowledgment
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W. D. Stuart for giving him an opportunity to make this review work for this conference. He is grateful to M. Ohtake for his careful reading of the manuscript, M. Imoto for permitting him to use his figure before publication, and L. Gao for sending him papers on coda study in China. In addition, he would like to thank A. V. Nikolaev, with whom the idea of the coda instability in the stress accumulation stage was first brought up in discussion.

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xWoSo

[ESIS]= erg/km 3, |V^]=erg . |gj=km" , [rj=|fit]-km

Figure 1. Reference space-time distributions of the single isotropic scattering model for coda energy density without coda attenuation (Qc~*=0): W 0, S wave source energy; go, total scattering coefficient; f, frequency; r 0, hypocentral distance; t, lapse time; ft 0, average S wave velocity [after Sato, 1977].

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Precursory tilt change
in July 21-31. 1983

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ENZ» ...;:

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'SMB VY

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The eastern Yamanashi earthquake

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Ms=6.0

0 -30km
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Depth

-30km

Figure 2. Results of two different kinds of analyses can be explained by the decrease of Q s -1 in the hemisphere (speckled region) after the eastern Yamanashi earthquake [after Sato, 1986a].

414

logtf-o

TRU

M

Open : Oct. 1, 1981 - Aug. 7, 1983
Closed : Aug. 8 - 31, 1983
Cross : Sep. 1, 1983 - Apr. 30, 1984

Figure 3. Plots of the logarithm of coda duration (s) at TRU against average magnitude [after Sato, 1986a]. Coda durations before the mainshock (open circles) are shorter than those after
it (closed circles and crosses) for earthquakes of the same magnitude. A broken line indicates the linear regression.

415

EASTERN YAMANASHI

P1

0.4-

0.2_

0.4.

MS6.0

60km

40km

1980 1981 1982 1983 1984

Figure 4. Temporal variations of the ratio Pi for the number of earthquakes of 2<M L <2.5 to that of 1.5£M L <2 within a radius of 40km and 60km from the mainshock hypocenter [After Imoto and Ishiguro, 1986]. High Pi values correspond to low b values.

416

Western Nagano Earthquake of Sep. 14,1984
Feb. 1982- Dec. 1984

GER

35.807* N
% *.
""VXMalnshock

(Ms- 6.8) Kamo-Tokal Network of NRCDP
37"N

139* I4I°E

(a) (b)

Figure 5 (a) Epicenter distribution before and after the western Nagano earthquake, Japan. Earthquakes which took place in the
core volume are used for the coda analysis at GER.
(b) Distribution of seismic stations and the mainshock epicenter (star) [After Sato, 1986c].

417

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"? 1.4

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1.2

1.6 1.8 2.0 2.2 2.4

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Figure 6. Plots of the logarithm of coda duration (s) at GER against average magnitude: crosses, period A (Feb. 18, 1982~May 7, 1983); open circles, period B (May 7,1983~Sep. 21, 1984); closed circles, period C (Sep. 22~Dec. 30, 1984); a solid line, linear regression [after Sato, 1986c].

418

Western Nagano Earthquake (M s = 6.8)

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Number

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Figure 7. Temporal variations of the residual of the logarithm of coda duration from the linear regression before and after the western Nagano earthquake. Here, a bold line indicates the mean
in each period.

419

Radon concentration

. Nagano E

1980 1981 1982 1983 1984

Figure 8. Weekly observation of radon concentration in soil gas around the epicenter of the western Nagano earthquake, Japan by using cellulose nitrate film: A, epicentral distance; T.D., a track density; MA, moving average for 5 weeks [after Katoh et
al. t 1986].

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FOCAL AREA

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Figure 9. Temporal variation in the slope of the pulse width of initial P wave against travel time at GER for the earthquakes that occurred in the aftershock region: P, Feb.l, 1982- Sep. 14, 1984;
A-I Sep. 14-15, 1984; A-II, Sep. 15- Nov. 13, 1984; A-III, Nov.
14, 1984- Sep. 13, 1985; A-IV, Sep.14- Dec. 31, 1985. Ext.-Area,
Feb. 1, 1982- Dec. 31, 1985, the exterior of the aftershock region. The shaded region indicates the 90X confidence interval [After Ohtake, 1986].

421

(a)

(b)

(c)

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TwT Dec .151971 ft

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March 23-24,1978

1974 M975 T 1976 ' 1977 ' 1978 M979 ' years

13 1963

1959 H960 M961 ' 1962 ' 1963 ' 1964 ' 1965

Figure 12. (a) Temporal variations in parameter a (s" 1) averaged over all channels at three stations surrounding the aftershock area before and after the Ust-Kamchatsk earthquake, where a is the anomaly of the logarithmic steepness of coda envelope,
(b) The Iturup earthquake (M s =8.0). (c) The Urup earthquake (Ms=8.1). [After Gusev and Lemzikov, 1985].

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(a)

- 2.0

a -2.2

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(b)

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Figure 15. (a) Temporal variations of log(Q c ~ 1 f) at 5Hz band, where a bold line and sand area show the mean and the standard deviation, respectively, in each period; crosses, Feb. 22, 1979- Sep. 1, 1979; open boxes, Oct. 20, 1979~0ct. 15, 1982; closed boxes, Feb. 12, 1983- Sep. 5, 1984. (b) Monthly number of earthquakes of K>7 which occurred in the aftershock region. [After Sato et al. , 1986] .

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300

8 20°
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OCT.31,1983
M6.2 5Hz

1 20 40 60
AUG. OCT MOV DEC.
1976 1983 1984 SEQUENTIAL EVENT NUMBER

80

Figure 17. Temporal variation of Q c at 5Hz band before and after the Misasa earthquake, Japan [after Tsukuda, 1985].

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7.12 7. 14.
6.25 7-3 7.4- 7.S 7.6 7.7 7.97.1 07 . I I 7. 137. 15 TIME

Figure 19. Temporal variation of the ratio of coda duration of horizontal component to that of vertical component (TH/TZ) before and after the Jianchuan earthquake in Yunnan, China [after Van and Mo, 1984].

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VII 1978
V 1979
ffi 1980
X~XE 1980
IX 1981

Figure 20. Temporal variation in the coda power spectrum of horizontal component explosion seismograms in the radial direction for lapse time of 21~40s observed at Shaambari. The shot point was the Iskanderkuli Lake in Soviet Central Asia [after Gamburtseva et al. , 1983].

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Figure 21. Temporal variations in coda parameter of explosion seismograms before and after the Liyang earthquake in China, (a) Configuration of the shot point (Ma'anshan mine), focal area, and two stations. (b) Temporal change in k (s~ 2), where k is the negative gradient of the predominant frequency f of coda against lapse time t: f=a-kt. (c) Temporal change in the exponential decay parameter a (s~M of coda envelopes. [After Gu et al., 1982].

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Lapse time

40 50 60 sec

(a)

Figure 22 Synthesized envelopes of three component seismograms of a M L =3 earthquake at a distance of 30krn in the inhomogeneous lithosphere based on the single scattering approximation for vector waves. The randomness of elastic moduli is characterized
by the exponential auto-correlation function with 10% in RMS fractional fluctuation and a correlation distance of 2km. The normal vector n and the slip vector s characterize the point shear dislocation. (a), (b) and (c) show different focal mechanisms. The third axis is taken to the radial direction.

[After Sato, 1984a].

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Fig. 24 (a)

Figure 24. Fracture experiment of a big granite block under the uniaxial loading [after Sobolev, 1984]. (a) Appearance of clusters of AE hypocenters before the fracture. (b) Temporal variations of V P along various ray traces against loading stress.
The scatter of V P increased before the fracture.

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Stress
(bar)
0 SO 100 150

.c *->
o.

Q

Figure 25. Horizontal in-situ stress (SH) in a borehole at Ashikawa-rnura, Japan measured by the hydraulic fracturing experiment [after Ikeda and Tsukahara, 1986].

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Preliminary Results on the Temporal Stability of Coda Waves in Central California from High-Precision Measurements of Characteristic Earthquakes

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Abstract

Characteristic earthquakes are used to study the temporal stability of amplitudes and travel times along the San Andreas and Calaveras fault systems in central California. The essentially perfect replication of the waveforms for each member of a characteristic earthquake family permit us to make direct comparisons between seismograms of related events at a common station. Wood-Anderson seismograms from Mt. Hamilton for M ~ 4 events from the Stone Canyon region along the San Andreas fault show that the ratio of the maximum shear wave amplitude to the coda amplitude varies by as much as 60 percent within a given-family in the time period from 1934 to 1984. The pattern of changes do not have any obvious temporal association with M>5.5 earthquakes occurring in the region between source and receiver. Travel time measurements of coda waves for microearthquakes (M 1.5 to 2.0) located near the 1979 M 5.9 Coyote Lake earthquake reveal that a small, but significant, travel time delay occurred within a limited region at about the time of the earthquake. If the delayed waves are shear waves, a velocity change of about -0.2 percent in a limited region near the southern end of the rupture is implied.

Introduction

The possibility that the seismic properties of the earth's crust undergo premonitory changes (see Sato, this volume, for a review) is a very hopeful one for intermediate-term earthquake prediction because they are among the most straightforward quantities to measure with high precision. In principle, all that is required is a fixed source-receiver geometry and a repeatable source. High precision measurements of travel times for both direct and reflected phases have been successfully made for nearly a decade by Clymer and McEvilly (1981), who used a mechanical vibrator as the energy source and achieved a precision of about 1 ms. Poupinet, et al. (1984) demonstrated how repeating microearthquakes could also be used to measure travel times to a comparable precision. Each method has its advantages: controlled sources can be used to make measurements at will; natural sources directly sample seismogenic depths and may be used to extend the record well into the past. Indeed, for retrospective studies, repeating earthquakes may be our best source, as it now appears that precisely repeating earthquake sources are commonplace, if not ubiquitous (Ellsworth and Dietz, 1986).

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We report here on two studies where repeating earthquake sources, or characteristic earthquakes, are used to monitor the seismic state of the crust. In each study, it is the remarkable coherence of the waveforms from recurrence to recurrence that permits us to make high-precision measurements of attenuation and velocity using otherwise routine seismic data. In the first case we examine the amplitudes of S-waves and coda waves from characteristic earthquakes in the Stone Canyon, Calif., region along the San Andreas fault as recorded on the Wood-Anderson seismograph at Mt. Hamilton (A = 75 to 90 km). Travel paths from this area pass through or near the 1961 Cienega Winery earthquakes (ML 5.6, 5.5), the 1979 Coyote Lake earthquake (ML 5.9), the 1984 Morgan Hill earthquake (ML 6.2) and the 1986 Tres Pinos earthquake (ML 5.7) (Figure 1). In the second case, we examine coda waves from characteristic earthquake pairs (e.g. earthquake doublets) located on the Calaveras and San Andreas faults near Hollister, Calif. The eight pairs of events studied cover the time interval from June 1978 to May 1981, which includes the 1979 Coyote Lake earthquake.

Characteristic Earthquakes

The distinguishing trait of characteristic earthquakes is their recurrent rupture of the same fault segment in events of nearly equal size. The evidence for characteristic behavior, in which coseismic slip on an individual fault segment always falls within a narrow range, was first proposed on the basis of geologic investigations of large earthquakes by Schwartz and Coppersmith (1984) and by Sieh (1981) (uniform-earthquake model). This model has since been extended to earthquakes of smaller size on the basis of the highly correlative waveforms produced by apparently repeating earthquakes (e.g. Parkfield, Calif, earthquakes, Bakun and McEvilly, 1984).

Poupinet et al. (1984) recognized that characteristic earthquakes, sharing
a common location, magnitude and mechanism, provide a powerful tool for tracking temporal variations in the state of stress, since precisely repeating sources will yield identical seismograms as long as the source and medium remain invariant. Even subtle changes in either the source or medium will result in changes in the observed wavefield. Because seismograms for pairs of spatially distinct events will also appear to be similar at periods corresponding to wavelengths longer than the separation of their centroids (Geller and Mueller, 1980), high coherence alone does not necessarily imply a repeating source (e.g., Spieth, 1981; Pechmann and Kanamori, 1982). Care must be taken to separate artifacts created by displaced sources from those attributable to changes elsewhere in the earth.

The results to be described below are based upon two different suites of events that we believe represent characteristic earthquakes. The first suite includes moderate-sized events from the Stone Canyon region along the San Andreas fault occurring between 1934 and 1986. While we believe that these events form sets of characteristic earthquakes, this point has not yet been demonstrated for all the sets. The other suite of events comes from Frechet's (1985) exhaustive study of earthquake doublets. These doublets, M 1.5 to 2.0 have an average separation of their moment centroids of 17+_ 6 m. Unless stress drops for these events are in the kilobar range, they must involve repetitive strain release from the same location on the fault.

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To us, the existence of the characteristic earthquakes themselves,
spanning a range of 109 in seismic moment, is the most interesting result, and bears much further elaboration and consideration. However, our purpose here is focussed more narrowly on the issue of possible changes in the seismic properties of the earth as intermediate-term earthquake precursors.

Shear Wave and Coda Wave Amplitude Variations

Wood-Anderson seismograms from Mount Hamilton (MHC) provide an
uninterrupted instrumental record spanning more than half a century which we shall use to examine spatial and temporal variations in the wave propagation characteristics of the crust in central California. The key to this work is our discovery of a remarkable suite of characteristic earthquake sources along the San Andreas fault some 80 km to the south of MHC (Figure 1). The earthquakes group into unambiguous sets of events on the basis of their virtually identical seismograms. In all, we present results for 28 events, ML 3.7 to 4.9, grouped into 8 sets (Table 1). The events are distributed along a 20 km-long segment of the San Andreas fault zone, and have focal depths ranging between 4.6 and 11.3 km (Figure 2). These 8 sets of events, plus 3 mre not considered here, include half of the M 4+ events occurring in this region during the period from 1934 to the present. We are still in the process of identifying all of the characteristic sources for this area and anticipate that some of the presently unmatched events will also fall into families of repeating events.

An example of the seismograms used to define the sets (with particularly clear records) appears in figure 3. The phase coherence between these two seismograms is essentially perfect from the p-wave onset through the entire duration of the event. Details at frequencies as high as 2-3 hz are clearly common to both records. (Use of a transparency copy of one seismogram is suggested as one way to see the correlation.) Although the quality of these records is unusual in that the entire record can be clearly seen, we require the same high degree of correlation throughout the legible portion of each seismogram for every event assigned to a set. This interval invariably includes at least one minute of coda and usually includes some of the body wave arrivals as well.

To compare the amplitude characteristics within and between the sets of characteristic earthquakes we measure the amplitudes of precisely the same wavelet for each member of the set. For simplicity, measurements are taken at the shear wave maximum and at prominent peaks in the coda where the accuracy of the measurements is best. In effect, we are determining the peak envelope of the seismogram. If the seismograms are locally scaled versions of each other, then these measurements record the relative amplification of the seismogram at each measurement point and can be used to study changes in its shape. The null hypothesis would be that each record is simply a scalar multiple of the other members of the set. In this case, the scale factor is the relative seismic moment for each member of the set.

Before discussing possible time-varying properties of the records, we examine the average shapes of the seismograms. Amplitude measurements for set 1 (figure 4) indicate that there are strong similarities in the envelope of the seismogram for each member of the set. Only the first member of the set (370217) shows any significant departure from the average trend.

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To examine the similarities in the coda envelope between different sets of characteristic earthquakes we compare the averaged peak amplitude decay curve for each set in Figure 5. For convenience, we normalize each curve by the maximum S-wave amplitude. It is somewhat surprising to us that there is as large a variation in the coda envelope shape from source zone to source zone given the small solid angle subtended by the entire source region at the Mt. Hamilton station.

Specifically, relative amplitudes of the coda waves arriving 30 s after the S-wave and normalized by its maximum amplitude vary by up to a factor of
3. Since all events have focal mechanism solutions corresponding to right-slip on the San Andreas fault, it appears unlikely that the source radiation pattern is controlling the observed variation in amplitude behavior. Note also that there is no simple correlation between the decay rate of the coda and source position or depth (Table 1). Our result suggest that extreme caution must be used when comparing coda amplitude decay behavior for events with even slight differences in their locations.

Amplitude comparisons between the members of each set of characteristic earthquakes should, in principle, provide a more stable and reliable means for assessing time-dependent wave propagation characteristics than the comparison between unrelated events. The approach used here is to compare the shear wave

amplitude (A s) with the coda amplitude (Ac) as measured at a fixed time after the S-arrival. Variations in the ratio of these amplitudes between

members of a given set should be sensitive to changes in the wave propagation characteristics of the earth, among other factors. It should be noted that the results for each set of related events cannot be directly compared to other sets, since the intrinsic value of the ratio A S/AC is not fixed
(figure 5). What can be compared between families are the relative variations in the ratio A S/AC determined for each set with the absolute level for
each set as a free parameter. Thus, with the available data we can search for consistent, but non-unique temporal patterns.

As a first cut, we compare the ratios A S/AC with Ac measured at the
same position on each record within a given family. A? is always measured in the interval from 27 to 34 s after the maximum amplitude. The exact point of measurement depends on the arrival time of a clear peak on the seismogram for each component. The region sampled by a single scattered S-wave arriving in this time interval is approximately limited by the dashed ellipse in figure 1.

The results (figure 6) imply a minimum variation in A S/AC of at least
a factor of 1.6, which is slightly larger than our estimated extremal bound on the errors of +_ 20 percent. The trends defined by the different families are not jointly interpretable in a simple way; several possibilities which minimize the scatter of values between the families are shown in figure 6b-6d. If a common precursory change precedes the largest events occurring within this region, we have either failed to sample it or it is below our noise level.

A skeptic may, of course, argue that there is no variation in AS/AC,
and all the measured changes simply reflect noise in the measurements. Referring to figure 3, one can discern a clear relative change in this ratio for these two seismograms. While the origin of this amplitude ratio change may be in doubt, we feel that it cannot be dismissed as an artifact.

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Coda Wave Travel Time Variations

Poupinet, et al. (1984) applied frequency domain cross correlation methods to the analysis of a pair of events (here referred to as doublet C2) with a common centroid (^ 10m) and determined that while the body waves were virtually identical, the coda waves at several stations were delayed so that the seismogram from the second event appeared to be stretched by up to 0.5 percent (Figure 7). Although an extremely small change, the coherence of the waveforms made such a high-precision measurement readily achievable using standard FM-telemetry data.

The geographic distribution of retarded coda wave arrivals suggested that the shear-wave velocity dropped in a well-defined region between the time of the first and second events forming the earthquake doublet. As the source location, region of observed coda wave retardation and its timing were all plausibly related to the August 6, 1979 Coyote Lake earthquake (M^ 5.9),
they suggested that the observed variation in coda-wave arrival times was related to this earthquake, although other possible explanations were also advanced.

Frechet (1985) studied this possibility through the systematic analysis of other earthquake doublets in the same region. In all, he found seven additional doublets where the relative separation of the hypocentroids were no more than 30 ^ 10 m, and which most probably represented the repeated rupture of the same area on the fault plane. The remarkably high coherence between seismograms for each doublet requires that differences in orientation of the moment tensor for each source pair cannot exceed 5° (Frechet, 1985). It is this high coherence of the seismograms that made possible the precise determination of the phase of arrivals.

His results (Figure 8) confirmed the earlier observations of Poupinet et
al. The common factor linking the doublets displaying temporally varying coda-wave arrival times appeared to be the occurrence of the Coyote Lake earthquake. The variations do not correlate in any simple way with seasonal rainfall (figure 9). Water table variations also do not show any obvious correlation with the timing of the velocity change (see Fig. 14., Schulz et al., 1983). The shape of the region with changing velocity, as mapped by the position of the stations within the network, was found to depend somewhat on the location of the source. However, the most profound variations always seemed to involve stations situated immediately south of the aftershock zone of the Coyote Lake earthquake, principally HOR, HFH and HKR.

Since only relative changes in velocity could be measured, a unique interpretation of these patterns cannot be made. Here, we shall attempt to produce a simple description of the sequence of events by tracking the average behavior of stations HOR, HFH and HKR. Two interpretations (Figure 10), one which minimizes variations before the Coyote Lake earthquake and one which minimizes post-seismic variations, each imply a step-like decrease in shear-wave velocity of a few parts per thousand at about the time of the earthquake. A change of this magnitude is consonant with its origin in the stress redistribution of the earthquake (Poupinet, et al. 1984), although such a conjecture is clearly non-unique. We do, however, believe that the observed stretching of the coda cannot be entirely relegated to either an instrumental artifact or a source effect, and thus represents a clear example of a velocity change. If tectonic, it implies that coseismic variations of a few parts per thousand occur.

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These same doublets also yield a wealth of information on the stability of coda wave amplitudes spanning the time of the Coyote Lake earthquake. Dramatic changes in their decay rate apparently did not accompany the changes in velocity. However, this question deserves a more thorough treatment than the visual examination of records made thus far. As with the longer-term record of amplitude variations from Mt. Hamilton, the temporal sampling is insufficient to draw a firm conclusion regarding amplitude variations.

Discussion

The two studies described above illustrate the power and utility of characteristic earthquake sources as tools for monitoring the crust. These examples emphasize the measurement of path effects and, to the degree that they are successful, rely upon the nearly exact repetition of the source. One possible criticism of the interpretations of these data would be that they are artifacts created by subtle changes in the source, including its location, mean orientation and dynamic details.

While it is true that the analysis depend almost exclusively upon frequencies at or below the source corner frequency, systematic amplitude variations for doublets have been reported. Fremont (1984) found that the spectral ratio of both P- and S-waves observed at certain stations were not constant and changed by up to 3 db/octave at frequencies below the corner frequency of the events (3 to 15 hz). Changes for a pair of events occurring on the same day and separated by 66 m were more extreme and displayed an azimuthally dependent variation in the slope of the spectral ratios in the 3 to 15 hz band. These results have not been satisfactorily explained to date and may be related to source directivity effects. If so, it raises the concern that the amplitude variations apparent in Figure 3 or 6 may have a similar origin. Alternatively, these events may not rupture the same area on the fault plane, so that the slight differences in the paths could account for the observed differences. We think that this possibility is remote. Both broader band observations (than the Wood-Anderson seismograph) and data from other azimuths would seem to be required to resolve this question.

It is less obvious to us, however, that the time delays in coda wave propagation can be explained by the same mechanism. The majority of the events which show significant changes must represent recurrent strain release in precisely the same zone, as their moment centroids are virtually identical. Specifically, the centroid separation for doublets C2, S7 and A867 are 12 +_ 6, 12 _+ 9 and 4 +_ 9 m, respectively. The observed changes in the coda appear to be too large to be explained by the alternatives suggested above, or do not uniformly affect all stations within a given azimuth range. Both the spatial extent of the region with changing velocity and its timing suggest a connection with the Coyote Lake earthquake.

Acknowledgements

We wish to thank the Director and staff of the Seismographic Stations, University of California at Berkeley and the Director and staff of the Seismological Laboratory of the California Institute of Technology for their assistance and cooperation in making the Wood-Anderson seismograms available to us. Our thanks also go to the generations of seismologists at both of these institutions whose dedication to the collection and preservation of these remarkable records made this study possible. We also thank Bernard Chouet and Dave Oppenheimer for their reviews of the manuscript.

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Figure Captions

Figure 1. Location map of central California showing the source region
(hatched box) and the receiver (MHC) used in this study; M > 5.5
events within the study area (stars); and the maximum raypath
distance traveled by backscattered S-waves originating in the
source area and arriving at MHC at t-t s = 35 seconds (dashed
ellipse).

Figure 2. Distribution of characteristic earthquake sets within the source
area in (A) plan view and (B) longitudinal projection onto a
vertical plan paralleling the San Andreas Fault (AA 1). Average
magnitudes are given for those sets used in this study.
Individual events are listed in Table 1.

Figure 3. Seismograms from the EW component of the Mt. Hamilton
Wood-Anderson instrument for the two most recent events in
characteristic source Set 9. The high coherence between the two
records can be easily seen by duplicating this figure on clear
plastic paper (overhead transparency) and overlaying one record
with the other.

Figure 4. Amplitude decay traces for individual events of Set 1. Open symbols represent maximum S-wave amplitudes for seismograms
wnere the actual peak-to-peak maximum was not measureable, but
was determined by the relative ratios of alternate S-packet
waveforms. Dashed line connects coda amplitude measurement
across missing data.

Figure 5. Peak amplitude decay curve for each characteristic earthquake
set. Dashed line connects amplitude measurements across poorly
sampled interval. The maximum S-wave amplitude (at t=0) is set
to the same arbitrary value for each trace.

Figure 6. Ratio of relative maximum S-wave amplitude to the relative coda
amplitude measured 27 to 34 seconds after the S arrival. These
ratios were determined from either the NS, EW, or both components of the Mt. Hamilton Wood-Anderson depending on the
quality of the records for each set. Dashed lines denote time
of M > 5.5 earthquakes located along a 110 km, NW-SE trending
zone having the source area at its southern end and MHC near its northern end (See Figure 1 for locations). Only relative variations within each set can be measured; therefore no unique interpretation of the patterns is possible.

A. Minimizes the overall scatter among sets.

B. Equates the ratios of the events occurring after the 1979 Coyote Lake event.

C. Equates the ratios of the events occurring between the 1961 Cienega Winery events and the 1979 Coyte Lake event.

D. Minimizes scatter among the earliest events of each set.

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Figure 7. Superposition of seismograms at station HORV for doublet C2. The seismogram for the second event (820228052) is delayed by up
to 0.2 sec 38 sec into the record, corresponding to a velocity
change of -0.5 (Reproduced from Frechet, 1985).

Figure 8a. Relative S-wave velocity variations (units of 0.1) for each of the following doublets:
C2: 780615 1248 and 800228 1231 (top left)
S7: 780828 0533 and 810523 0026 (top right)
A8 67: 780905 0814 and 790827 1854 (bottom left)
A7 13: 790114 0235 and 801126 1005 (bottom right)
Solid circle denotes the location of the doublet, dashed lines
enclose areas where A V s > o.l dotted lines enclose areas

where A Y s < -o.l . Epicenter (star) and aftershock zone
(hatched area) of the August 6, 1979 Coyote Lake earthquake are
shown for doublets which span the event.

Figure 8b. Relative S-wave velocity variations (units of 0.1) for each of the following doublets:
S4: 790609 0031 and 810615 1429 (top left)
A6 34: 790806 2045 and 790908 0122 (top right)
A7 21: 790911 0357 and 801126 1005 (bottom left)
A6 57: 800331 0143 and 800901 1610 (bottom right)

Figure 9. (A) Rainfall in the Hollister region 1978 through 1981. (B) line segments representing the time spanned by each doublet (see
figure 8 for event pair dates and times).

Figure 10. Two interpretations of temporal variations in shear wave velocity deduced from observed changes in coda wave arrivals.
Doublets are identified by letters. The half-filled point
labelled A721 belongs to an event triplet which also includes
A713. The point labelled A713 may thus be compared to either
point labelled A721. (Top) Velocity variation obtained by
assuming constant pre-earthquake velocities. (Bottom) Velocity

variation obtained by minimizing co-seismic and post-seismic variations.

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VOLUME II

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NATIONAL EARTHQUAKE PREDICTION AND HAZARDS PROGRAMS

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MENLO PARK, CALIFORNIA

1987

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Precursory Seismic Quiescence: Past, Present, and Future

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ABSTRACT

Precursory seismic quiescence has played a major role in most of the successful earthquake predictions that have been made to date. In addition to these successes, the number of detailed post-mainshock documentations of precursory quiescence is steadily growing. These facts suggest that precursory quiescence will play an important role in earthquake prediction programs of the future. For this reason it is important to critically evaluate the present state of knowledge concerning this phenomenon.

The history of observations of seismic quiescence in the rupture zones of upcoming mainshocks extends back to some of the earliest studies of seismic precursors. Many of these studies mixed what we presently call seismic quiescence with the observation of seismic gaps although now the differences between these phenomena are understood. The next papers examined "seismic preconditions" and were really studies of spatial variations in seismicity rather than studies related to intermediate-term earthquake prediction. The first real studies of temporal quiescence appeared between the late 1970's and early 1980's. These papers demonstrated the importance of quantitative evaluation of seismicity rates and the benefits of systematic analysis.

During the early 1980's the impact of man-made effects on seismicity rates was demonstrated for the first time. A number of important studies have demonstrated this impact on large and small events in teleseismic and local catalogs. The early 1980's have been a period of "catalog awareness" in seismicity studies which has led to substantial and important advances in understanding these data. In addition, work is presently underway which will result in much improved catalogs which can serve as the basis for earthquake prediction studies of the future.

Despite this progress in catalog understanding, the identification and correction of man-made seismicity changes remains as the major barrier to earthquake prediction using these data. Effects of man-made changes are apparent in many past studies of seismicity patterns yet have gone unnoticed by the investigators involved. The results of many of these studies are, therefore, difficult to evaluate.

Recent experience with real-time anomalies has demonstrated the necessity
of determining the false alarm rates associated with quiescence precursors. Several investigators have presented systematic approaches to determining these rates, and these techniques have now been applied to a number of cases. These indicate that false alarm rates may be very low (i.e. < 10%).

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Determination of false alarm rates depends on quantitative definitions of anomalies and statistical evaluations of their significance. Several techniques have been used to evaluate this significance and most give similar results. This suggests that the statistical framework of the studies is not as important as careful data treatment.

There have been a number of successful predictions which have been made
on the basis of seismic quiescence and a number of presently open predictions. All of these past cases provide important lessons for present and future work.

There are many presently unanswered questions regarding seismic quiescence which must be answered before we can determine the reliability of this phenomena as a precursor. Steps that must be taken include: Systematic analysis of existing seismicity data using a wide variety of statistical tools and development and implementation of real-time anomaly recognition programs. In addition we must agree on standard techniques for describing and testing possible anomalies so that the evaluation of these cases can be done consistently and without bias. These standards must include techniques for demonstrating that the anomaly is not caused by changes in the network and for determining false alarm rates. Finally, we need to integrate studies of other geophysical parameters and tectonics with seismicity studies.

INTRODUCTION

Several recent successful predictions and studies of precursors to past events suggest that seismicity patterns can play an important role in earthquake prediction programs of the future. A number of authors have examined seismicity prior to past earthquakes and proposed various patterns as precursors. Kanamori (1981) and McNally (1982) review many of those studies and outline the basic types of seismicity precursors (foreshocks, quiescence, swarms, and doughnuts). Of these, seismic quiescence has received the most attention recently and, therefore, it is the subject of this review.

This paper discusses several aspects of the problem of precursory quiescence. First, I describe briefly the historical foundation of our present understanding of this phenomena. Second, I discuss several important problems which one faces in studies of seismicity patterns and demonstrate the effects of those problems on past studies by myself and others. I then briefly review successful predictions which have been made on the basis of seismic quiescence and predictions which are presently in effect. Finally I outline the steps that must be taken in the future if seismicity patterns are to fulfill their potential in earthquake prediction studies.

Why Seismicity Data?

Seismicity data are important for earthquake prediction studies for a number of reasons. First, they are the most abundant data available. Hundreds of thousands of earthquakes have been recorded and cataloged from all seismically active parts of the world. These catalogs cover time periods which extend from tens to thousands of years and, of course, their quality and

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completeness vary considerably. This large database is important for a number of reasons, primarily it makes it possible to study seismicity patterns prior to a number of mainshocks from different tectonic regimes. This will be crucial for understanding the complexity associated with precursory changes in seismicity patterns as well as for developing empirical guidelines for precursor time estimates and the associated errors. In addition, this database provides the information which will be needed to determine false alarm rates associated with seismicity precursors.

The second reason that seismicity data are important is that they are the only data which come directly from the crustal depths where mainshocks occur, so they directly reflect processes which are occurring in the regions of mainshock preparation. Earthquakes occur when the stress overcomes the strength on some section of fault, so they give us important information about changing stress or strength on the fault surface.

Seismicity also provides information which can be remotely sensed, allowing one to monitor large regions with a relatively small number of instruments. This unique characteristic makes seismicity data a prime candidate for the first level of any precursor recognition and monitoring program. The large data base mentioned above can be used to test and develop automatic systems for carrying out this monitoring.

The final reason for including seismicity patterns in an earthquake prediction research program is that they have been the most successful earthquake precursor examined to date. All of the successful predictions that have been made have relied to a great extent on analysis of seismicity data. These successes clearly indicate the necessity of expanding our studies of seismicity patterns, not only to learn what we can from past successes, but to lay the groundwork for future predictions.

QUIESCENCE STUDIES RELATED TO EARTHQUAKE PREDICTION

Mogi laid the foundation for many of the modern studies of seismicity patterns and earthquake prediction in his series of classic papers published during the 1960's. He developed the fracture theory of earthquakes based on the application of laboratory observations to seismicity patterns around Japan. The phenomena he studied included seismic gaps, migration, b-values, as well as temporal and spatial distributions of foreshocks and aftershocks. His papers contain many observations that remain important today.

Most of Mogi's work was limited to study of events with magnitudes _> 6 because of the catalogs that were available to him. Many more recent studies have not considered this when comparing Mogi "doughnuts" with observations of quiescence in smaller events. The physics behind these two observations are probably quite different and they should not be confused.

Mogi looked for the occurrence of foreshocks (at the M>^ 6 level) prior to
a number of events in Japan because such events were expected on the basis of laboratory experiments. He was surprised not to find these events in most cases (Mogi, 1969):

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Until now, the increase of seismic activity before a great earthquake has been supposed by many investigators.
In a few cases, great earthquakes were certainly preceded
by some foreshocks. In most cases, however, any marked
foreshock activity could not be observed."

The observations of changes in seismicity rates for smaller events made
by Mogi were presented in a discussion of the Tokachi-oki earthquake of May 16, 1968:

In this case, precise seismic data including smaller
events are available and so a detailed investigation of
the seismic activity before the great earthquake can be
made..... an appreciable increase of the seismic activity just before the great earthquake is not seen, but the activity has rather decreased gradually since 1961, except for 1965 in which two major earthquakes (M>6)
occurred in the focal region. This result indicates that
the foreshock activity in strict sense did not occur
before this great earthquake.

Here again, the expectation of foreshock activity is apparent, and the observation of quiescence is surprising.

Of the many fundamental observations which Mogi made, the most important
one for intermediate-term prediction is that, in the general case, there is relative quiescence prior to great earthquakes near Japan (at least at the M > 6 level). As Mogi points out, this is an unexpected observation, and one that cannot be easily explained in terms of laboratory observations. Mogi called on increased strength of the focal region to explain the lack of seismic activity, but he did not explain how this occurred during the preparation process.

Several papers applied the type of analysis used by Mogi to other areas. These included work in California by Wesson and Ellsworth (1973) and work in a number of regions of the world by Kelleher and Savino (1975). These papers were not really about earthquake prediction because they lacked recognition of temporal changes in seismicity patterns (Kelleher and Savino):

Although the pattern of spatial distributions of prior
seismicity is reasonably consistent, temporal changes in
prior seismicity are more difficult to analyze; the data presented in this study are insufficient to conclude that
any precise time interval before large earthquakes is associated with distinct seismicity changes.

These studies examined what Kelleher and Savino called seismic preconditions:

the magnitude, frequency, spatial distribution, and, where it can be determined, the temporal distribution of observed seismic events originating in and near the pending rupture zone for as many as several decades

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before the time of the mainshock.

In a modern context these papers are essentially studies of the relationship between spatial variations in seismicity and moderate to great earthquakes and, as such, they make several important contributions. Kelleher and Savino pointed out differences in behavior for different parts of the rupture zone. They concluded that most of the rupture zones of large earthquakes were quiet during the period prior to the mainshocks, however:

An important question remains unanswered: it is not clear
whether a zone at some time becomes quiescent as part of
a process premonitory to a large earthquake or whether
after the aftershocks of a large earthquake diminish, the
zone remains quiescent until the time of the next large
earthquake. Although examples can be produced that
support either alternative, the totality of data
available at present does not discriminate between these
alternatives.

Wesson and Ellsworth concluded that moderate events initiated in segments
of faults which were marked by a temporally persistent high level of activity. They formulated a conceptual model of fault behavior that stressed the importance of aseismic slip and included patches in which motion was inhibited because of differences in fault properties. These zones were zones of stress concentration and initiation of moderate earthquakes. This idea has become known as the asperity model for fault processes and the concept that asperities are important in controlling earthquake precursors as well as source properties has become almost universally accepted.

The first study of temporal changes in seismicity rates related to earthquake prediction was done by Ohtake and others and published in two papers (Ohtake et al., 1977 a,b). They examined eleven events in Mexico and Central America and concluded that most were preceded by a decrease in activity followed by an increase shortly before the mainshock. On the basis of these observations and the observation of a similar pattern in the Oaxaca region, they predicted that an event would occur in that region in the near future (Ohtake et al. 1977a). The predicted magnitude (7.5 +/- .25) as well as the location were determined on the basis of the size of the gap between the 1965 and 1968 Oaxaca events, rather than on the size of the anomalous area. No prediction of the time of the event was made, however on the basis of other cases in the region it is clear that the event was expected soon.

The Oaxaca event which occurred on November 29, 1978 matched the predicted event quite well (Ohtake et al., 1981). This successful prediction was a clear indication of the potential of seismicity rate studies for earthquake prediction.

An important innovation made by Ohtake et al. (1977b) was the use of computer generated space-time density plots for systematically examining seismicity rate variations in large regions. The use of some such techniques is clearly necessary for monitoring regions for seismicity precursors and the development of real-time tools for carrying out this monitoring remains an important task.

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The next set of papers which examined the problem of seismicity rates as intermediate-term precursors was published in the Proceedings of the Ewing Symposium on Earthquake Prediction. Kanamori (1981) reviewed the progress that had been made up to that time, Habermann (1981a) reported on a systematic study of seismicity rates in a number of seismic gaps, and McNally (1981) reanalyzed the data from Mexico and Central America. In addition, several predictions were made on the basis of presently occurring quiescence in Greece and in Mexico (Wyss and Baer, 1981; McNally, 1981).

The papers in the Ewing volume demonstrate a fundamental change in approach to earthquake prediction using seismicity patterns which occurred during the early 1980's. Kanamori (1981) recognized this change and described the two approaches:

In the [historical] approach, various seismicity patterns
are used as a clue to the physical mechanism of the
earthquake process.... In this approach, rigorous definition of seismicity patterns and uniformity of catalogs are less important than in the other approach. Although seismicity patterns thus somewhat loosely defined may not be used directly for prediction purposes, they can be used to identify a possible physical mechanism.... In the [more modern] approach, various seismicity patterns are treated as rigorously as possible in a statistical framework to establish an empirical algorithm for earthquake prediction. Essential to this approach are the uniformity of earthquake catalogs and rigorous definition of the seismicity patterns.

As Kanamori states, his paper continued along the same lines as much of the previous work. He showed space-time diagrams for a number of subduction zones and visually picked apparent periods of quiescence prior to a number of mainshocks. The paper included no quantitative determinations of background rates, no statistical evaluation of the potential anomalies and no analysis of possible problems with the data that might affect the conclusions. As we shall see below, this approach can easily lead to misinterpretations and results which are impossible to evaluate reasonably.

The more modern approach usually presents the data as cumulative numbers
of events as a function of time and relies on quantitative comparisons of seismicity rates. This approach was used by McNally (1981) and Habermann (1981a). McNally presented cumulative number curves for the rupture zones of large events in Central America and Mexico, visually picked anomaly onset times and evaluated the significance of the anomalies using an assumption that the seismicity was a stationary Poisson process. The same test was used by Ohtake et al. (1981) to evaluate the Oaxaca precursor, although they relied on space-time and magnitude-time plots for data presentation.

Habermann also used cumulative number of event plots for data presentation and proposed several statistical algorithms for objectively determining the anomaly onset times and for evaluating the temporal and spatial uniqueness of the anomalies. This approach is fundamentally different

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than that of McNally who used quantitative tools for determining the significance of visually identified anomalies. Habermann directly addressed the important question of false alarm recognition by using statistical tools directly for anomaly identification. In addition, he demonstrated that detection decreases in the teleseismic catalogs caused changes in seismicity rates that could easily be mistaken for precursors.

The papers by McNally and Habermann set the stage for the developments which have occurred since the time of the Ewing Symposium. They raised several important issues: First, they demonstrated the desirability of using quantitative tools for anomaly evaluation. This was a major step forward from the older techniques of seismicity maps, space-time plots and magnitude-time plots. Second, the specter of man-made changes in seismicity was raised for the first time. Finally, the problem of determination of false alarm rates was addressed.

Subsequent studies have shown that the differences between the statistical approaches to determining the significance of anomalies proposed by Habermann and McNally are minor, both produce consistent estimates. However, correction of catalogs for man-made effects and determination of false alarm rates remain today as the most serious barriers to earthquake prediction using seismicity patterns. We summarize recent work relevant to these problems in the next section.

Summary

The early work described above clearly suggested that large portions of rupture zones were quiet prior to mainshocks, but that the region of an upcoming hypocenter remained active. Note that 1) most of these studies did not determine when the rupture zone became quiet because of catalog limitations and 2) that the magnitude cutoffs used in many of these studies were fairly high, precluding direct comparison with more modern work. The observation of heterogeneous behavior within the rupture zone contrasts strongly with the ideas held during the reign of the dilatancy-diffusion model (the early 1970's), when fairly uniform precursors were expected throughout regions three times the size of the upcoming rupture zone.

Recently, more detailed analysis of precursors to several events in Hawaii (Wyss et al., 1981; Wyss, 1986), in the New Hebrides (Habermann and Wyss, 1984), and in California (Wesson and Ellsworth, 1973; Wyss and Habermann, 1986) have demonstrated that spatial heterogeneity also exists in the patterns of occurrence of smaller earthquakes in the rupture zones of upcoming earthquakes. The most noteworthy case is the precursor to the 1975 Kalapana, Hawaii event. The precursor to the event, and the source rupture itself (Harvey and Wyss, 1986), clearly tie the variations in observed seismicity precursors to spatial variations in stress (or strength) within the rupture zone. This complexity of observed seismicity rate precursors may make it difficult to interpret observed seismicity rate variations in real-time unambiguously even with perfect seismicity observations (e.g. Wyss and Burford, 1985).

Despite the ubiquity of precursory quiescence, there was little

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understanding of the mechanism which caused the observations. Mogi (1969) and Kanaraori (1981) both suggested that spatial variations in strength might explain the observations, though their envisioned relationships of the strength variations to the rupture were somewhat different. These explanations suffered from difficulty in explaining the onset of quiescence. If the focal area or the asperities were stronger, why did they experience background seismicity and why did the rate of seismicity in them change?

The importance of spatial variations in fault properties in explaining seismicity patterns, suggested by Wesson and Ellsworth (1973), was more fully developed and given some observational support in work related to the Kalapana, Hawaii event summarized by Wyss et al. (1981). They suggested that the quiescence was caused by aseismic slip on the non-asperity sections of the fault and that the asperities showed constant or increasing seismicity rates because of stress concentration.

UNSOLVED PROBLEMS

The most insidious problem facing scientists interested in using seismicity data for earthquake prediction studies is the pervasive occurrence of man-made changes in seismicity catalogs. It is a problem which exists in all studies and just doesn't go away. The omnipresent nature of the problem has been demonstrated by North (1977), Habermann (1982), and Perez and Scholz (1984), among others. Despite the rather clear results of these studies, most of the earthquake prediction community seems ignorant of the impact of the problem on earthquake prediction studies. For this reason we demonstrate the effects of man-made changes in a number of cases from the literature below.

If one can eliminate man-made changes as an explanation for an observed precursor, then the major problem which remains is the determination of the false alarm rate associated with the precursor. Few studies have addressed the problem of false alarm rates associated with seismicity rate changes. He outline an approach to the problem which we have developed and applied in several teleseismic and local cases.

The final problem which remains is the determination of the optimum statistical framework for studies of changes in seismicity rates. Our analysis indicates that this is not an important problem. Several statistical approaches have been proposed, and all seem to produce consistent results. Nonetheless, we outline those approaches here.

MAN-MADE CHANGES IN SEISMICITY CATALOGS

We have recently made great progress in understanding the impact of changes in networks and procedures on seismicity catalogs. Two general types of changes have been recognized, detection changes and systematic changes in magnitude estimates. A summary of techniques for recognizing and dealing with these changes has recently been published (Habermann, 1986b). The results of this work clearly indicate the importance of accounting for these changes in studies of seismicity rates for earthquake prediction. In this section we summarize some of the important results and demonstrate the impact of

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neglecting these changes on past studies.

Detection Changes

Most people involved in studies of seismicity precursors agree that the precursors are restricted to regions comparable in size and location to the rupture zone of the upcoming mainshock. If one finds a significant rate decrease in some region, therefore, it might be reasonable to expect that an event may occur in the near future in that region. The curve labeled ALL in Figure 1 shows the cumulative number of events as a function of time for the entire world between 1963 and 1979. This region clearly shows a strong decrease in rate during the late 1960's which is significant at the 99X level according to several statistical tests. No reasonable seismologist would propose, however, that an event would rupture all of the seismically active regions of the world (the region represented in this plot), so some alternative explanation must be sought for this rate decrease.

The other two curves in Figure 1 show that the change in rate is clearly restricted to the smaller events with the curve for larger events (4.5+) shoving a remarkably constant rate over the whole time period. The characteristics of the change during the late 1960's, the fact that it occurs in a large region and is restricted to the smaller events, are those that one would expect for a change which was caused by a decrease in detection. Habermann (1982a) presented these observations and proposed that the detection decrease was due to the closure of the VELA arrays (BMO, CPO, TFO, UBO, and WHO). The data in Figure 1 clearly show that this is a first order effect in the teleseismic catalogs, a fact pointed out earlier by North (1977). Subsequent regional studies have confirmed the global impact of the closure of these arrays (Habermann, 1982b, 1983, Wyss et al. 1983, 1984, Habermann et al. 1986, Habermann, 1986b).

The effect of the decrease in detection is very clear in the studies of seismicity rate changes in Mexico and Central America. Figure 2A shows the magnitude-time plot used by Ohtake et al. (1981) to demonstrate the Oaxaca precursor. The absence of activity prior to the 1978 event is clear in this Figure. Careful examination of the Figure, however, reveals a profound change in the seismicity in this region during the late 1960's. Notice the complete lack of smaller events after 1968. This change is certainly as strong as the change during 1975 for the smaller events, yet it was not recognized at the time of the prediction of the Oaxaca event or mentioned in the evaluation of that prediction presented during 1981.

The decrease in the number of smaller events is less obvious in the time-space plot for this region presented by Kanamori (1981) and shown as Figure 2B. In this case the symbol sizes are scaled to the magnitude so the detection change appears as a lack of small symbols after 1968. The fact that the change exists in all of Central America and Mexico can be visually confirmed in this presentation (with some difficulty).

Cumulative number plots for the region make the rate decrease caused by
the detection change more obvious than either of these other presentations. Figure 2C shows the cumulative number of events in all of Central America and

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Mexico between 1963 and 1979 for a number of magnitude bands. This Figure clearly demonstrates that the nature of observed rate changes is a strong function of the size of the events considered. The change in rate during the late 1960's occurred over a large region and was restricted to the small events. The effect of the decrease is, however, clearly observable in the set of all events because most of the events are small.

The final frame of Figure 2 shows the z-value which results from comparing the seismicity rates between February 1964 and November 1967 to those between November 1967 and July 1973 for all of Central America and Mexico (diamonds). Such plots are termed magnitude signatures and have been described in detail by Habermann 1983 and 1987. The plot shows the significance of the change in Central America and Mexico during 1967 as a function of magnitude cutoff. Data sets with upper cutoffs are shown on the left side of the plot and those with lower cutoffs are shown on the right side. This division is made because detection changes will effect the data sets on the left side of the plot more than those on the right so the effects of such changes are easy to identify in this presentation.

Figure 2D shows a clear picture of what the rate change during 1968 looks like. It is clearly strong in the smaller events and weak in the events larger than 4.6. Once again, these are the characteristics expected for a decrease in detection. Figure 2D also shows comparisons for nearly the same time periods from the Kuriles (boxes). The similarity of the pattern for these two distant regions confirms the global distribution of this change. Similar patterns have also been found in the Aleutians (Habermann, 1983), the New Hebrides (Wyss et al., 1983), and Tonga (Wyss et al., 1984).

This example demonstrates that seismicity rate changes can be a strong function of magnitude and that the distribution of observed changes in the magnitude domain can be very useful in determining the causes of the changes. The fact that this detection change occurred in Mexico has a considerable effect on how one evaluates the precursor to the Oaxaca event. If one extends the background period back to 1963, the estimate of the background rate will be too high, unless one applies a magnitude cutoff to the data. This cutoff sharply decreases the number of events and the significance of the rate decrease prior to the 1978 event (Habermann, 1981a). The alternative is to use no magnitude cutoff, and only consider the period between 1968 and 1975 as a background period, the approach taken by McNally (1981). This decreases the length of the background period substantially, again making recognition of rate changes more uncertain. Similar problems arise in all studies of seismicity rate changes in teleseismic catalogs and seriously degrade the resolution of these data.

Detection increases can also occur and change apparent seismicity rates. This type of change has occurred in regions where seismic networks have improved or where special studies have been carried out. An example of the results of such a study are apparent in the seismicity of the Kuriles presented by Kanamori (1981), shown in Figure 3.

The most striking feature of this Figure is the dark vertical band during 1961 which suggests that that year is very active for small events relative to other periods. However, when one examines the catalog for 1961, it is obvious

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that it is very different than for this region during other times. There are a large number of events with magnitudes reported by Moscow (MOS) which are essentially absent at other times. Kanamori suggests that this is also a tine of increased activity at the 6+ level. The time period involved (1961) is not a time of routine reporting of MS by NEIS, so most MS estimates listed are single station reports. The apparent increase in activity at this level simply reflects the inclusion of MOS magnitudes in the catalog. It is not reasonable to compare these magnitudes to those during other times without demonstrating that they are consistent with estimates from other stations.

Later in his paper Kanamori references the 1963 Kuriles event as an example of his conceptual model of precursory seismicity patterns. He states, "The 1963 Kuriles event....has a precursory swarm, quiescence, and foreshocks." Examination of his Figure 17A indicates that the quiescence is nothing other than the ambient level of activity and that the precursory swarm is simply the increased reporting during 1961. This example demonstrates how easy it is to be misled by relatively simple man-made changes in catalogs.

The last example shows that it is rather easy to mistake increased detection for increased activity. What about mistaking decreased detection for quiescence? Consider one of the observsations made in the papers in the Ewing volume, that for the Acapulco region, McNally (1981). The cumulative number of events curve for this region is shown in Figure 4. McNally uses a magnitude cutoff of 4.0 in this region. This is well below the level of completeness in the Oaxaca region (shown by Ohtake et al. 1981 to be 4.5) and below the magnitude of homogeneous reporting determined by Habermann (Figure 2D). Note the effects of the detection decrease during the late 1960's in Figure 4. Also note that the time of initiation of the anomaly, January 1977, is very close to the time of a clear decrease in world-wide reporting of smaller events (October, 1976, Figure 1). Without analysis of the distribution of the rate change in Figure 4 in the magnitude domain and of its spatial distribution, it is difficult to evaluate whether it is real or related to the generally low reporting during much of the quiet period. The facts that the detection decrease during the late 1960's is clearly visible in McNally's data and that the "precursor" begins at the same time as the worst period of reporting in the history of the PDE clearly suggest that the rate decrease may reflect a change in reporting rather than a change in the processes in the earth.

Misinterpretations of man-made changes as precursors are not limited to smaller events as in these two cases. Perez (1983) discusses the effect of a special effort by Gutenberg and Richter to locate as many events with M>6 as possible between 1930 and June, 1935 (Gutenberg and Richter, 1954, p.3). He points out that the rate of listing of events with 6.0 < M < 6.4 in the Gutenberg and Richter changes from 18 eqs/year between 1920-1920 to 60 eqfi/year during 1930-1934 and back to 17 eqs/year in 1936 to 1949:

It is obvious that the changes are due to a low level of detection and reporting capabilities for small events
prior to 1930, followed by the special effort of Gutenberg and Richter in 1930-1935, and then the world conflicts in 1936-48, and not to real changes in the
world seismicity rates..... Thus, if one analyzes the time

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distribution of events with M(GR) > 6 occurring since 1920 in seismic regions where great shocks occur decades later, and if a high level of activity is found in several regions in 1930-1935 followed thereafter by a lower level of seismicity, one could wrongly "find" a "precursory" burst of seismic activity in 1930-1935, or a period of "quiescence" preceding the mainshock. That has
actually been the case [e.g., Perez and Aggarwal, 1981,
Kanamori, 19811 if one uses the instrumental catalog as
is to study the seismicity before the 1952 Kamchatka,
1957 Aleutians, 1964 Alaska, and other very great
earthquakes. (Perez and Scholz, 1983).

Figure 5, from Perez (1983), clearly demonstrates the effect of the Gutenberg-Richter effort. The peaks during the early 1930's are just "artifacts of the heterogeneities of the catalogs", not precursors. The irony in this situation is that the great events were all nearly the same size, and the "precursor times" were all similar, so the special study by Gutenberg and Richter became the basis for many of the early precursor time - magnitude relationships.

Summary

The examples described above clearly indicate the impact of detection changes in teleseismic catalogs. They also demonstrate that even well documented detection changes can be mistaken for precursors. We showed several cases in which periods of increased reporting were identified as "precursory swarms" and one in which decreased reporting was proposed as a quiescence precursor to a future event. These cases are not unique, in fact finding nan-made changes which are proposed as precursors in the literature is surprisingly easy.

The effects of detection changes can generally be avoided by using a lower magnitude cutoff. Unfortunately, this can decrease the amount of data available considerably. Habermann (1983) discusses techniques for determining the lowest possible magnitude cutoff which allows one to keep the maximum amount of information without introducing detection changes into the data. Also, the examples presented here are limited to teleseismic cases, however, similar observations can be made in all seismicity catalogs we have examined including regional and local catalogs.

Systematic Changes in Magnitudes

As stated above, the effects of detection changes are rather easy to deal with, although the cost in terms of amount of data can be high. Recently another type of man-made change has been described which is much more difficult to deal with. This is a systematic change in magnitudes of events, usually caused by change in the station distribution used to determine the magnitudes.

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The first study which demonstrated the effect of systematic changes in magnitudes on earthquake prediction studies was the work of Perez and Scholz (1984) on the teleseismic catalog of this century. They made plots similar to Habermann's plot in Figure 1, but extended the time back to 1900 and considered only events larger than 7. These curves (Figure 6) clearly show strong changes in slope.

In all studies of systematic errors in magnitudes one needs standards to judge magnitudes against. Perez and Scholz established standards by assuming that the present reporting rate of the large events represent completeness and that the rate of occurrence of events with M > 7 on a global scale should be generally constant. With these assumptions in hand it becomes simple to adjust the magnitudes of events in the past in order to make the past rates equal to the present rates. They found that:

The MS of large events prior to 1908 have been systematically overestimated by at least 0.5 magnitude
unit, and the MS of events in the period 1908-1948 have
been consistently overestimated by 0.2 unit, relative to
the MS assigned to shocks occurring after 1948.

This study presented a bad omen for studies of rates of smaller events,
as one would expect the magnitudes of the largest events to be the most stable. If they were, in fact, unstable, then the magnitudes of smaller events may also have problems. North (1977) observed systematic changes in magnitudes of smaller events in the ISC catalog related to the closure of the VELA arrays discussed above. This occurred because these stations were very influential in teleseismic magnitudes of the 1960's and many had large magnitude biases.

Nishenko (1985) published the results of a study of seismicity rates in
the Valparaiso, Chile region done by Habermann. He showed that the 1971 Valparaiso event was preceded by 3.5 years of seismic quiescence, starting during 1968. The cumulative number of events as a function of time (CNET) curve for this region is shown in Figure 7A. Figure 7B shows data from a much larger (1700 km) segment of the Chilean subduction zone which includes the region shown in Figure 7A. The same period of decreased rates between 1968 and 1971 is apparent in these data. It seems unlikely that the precursor to an event with dimensions of several hundred km would extend over such a large region or that a decrease which was limited to the rupture zone would be so clear in the large surrounding region. The fact that it persists in the large region suggests that it may be due to a man-made change in reported seismicity.

Once again, the distribution of the change in the magnitude domain can
aid us in interpreting these observations. Magnitude signatures for the March, 1968 and July 1970 changes are shown in Figure 7C and D. The uppermost frames show the observed magnitude signatures. The general characteristics of these observed magnitude signatures deserve some comment as they shed quite a bit of light on the nature of these changes.

The change during 1968 has an important similarity to the magnitude signature shown in Figure 2 for the same time in Central America and the

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Kuriles (which we know to be related to a detection decrease). All three magnitude signatures show a platform of high z-values on the left side of the plot near the center. Our experience with this type of analysis indicates that this platform reflects a detection decrease. However, the left side of the plot for Chile lacks the clear peaks evident in Figure 2, as if the z-values on the far left side of the plot have been depressed. The right side of the plot for Chile is also different from those of Figure 2, showing strong rate decreases in nearly all bands. This suggests that the change in Chile is not as simple as the changes in Mexico or the Kuriles.

We have developed a technique of synthesizing complex rate changes in order to help understand what nay have caused then. The technique is described in detail by Habermann (1986, 1987). Essentially one takes the events in the background period and operates on then by applying a known magnitude shift or detection change. One then compares the modified set to the unmodified set by constructing a synthetic magnitude signature and then iterates the modifications until the best fit is achieved. This process is demonstrated in Figure 7C where a series of operations are applied to the data from the background period for the change during 1968 until agreement with the observed change is achieved.

The strong decreases which extend completely across the right side of the magnitude signature for the change during 1968 and the lack of a strong peak on the left suggest that the magnitudes decreased systematically at this time. This can be tested by examining a synthetic magnitude signature constructed by comparing the background period to itself with the magnitudes of all events decreased by 0.4. This magnitude signature is shown by the asterisks in the middle frame of Figure 7C. This synthetic fits the observations very well on the right side of the plot but shows strong increases when the data is viewed with an upper cutoff (on the left side of the plot).

Remember that the platform on the left side of the plot indicates a detection decrease, which is added to the synthetic by decreasing the number of events smaller than 4.6 by 50X to create the magnitude signature shown by diamonds in the middle frame of Figure 1C. Note that this detection decrease has little effect on the right side of the plot. With the detection decrease added to match the platform, it becomes clear that the previous magnitude shift is too strong in the smaller events, so the magnitudes of the events smaller than 4.3 are increased by 0.3. The final result simulates a decrease in magnitudes which is weaker for the smaller events. The final magnitude signature is shown in the bottom frame of Figure 7C and the fit between the observed and expected is very good. The final picture of this change during 1968 is then, a decrease in magnitudes combined with a detection decrease.

The change during July 1970 is simple compared to that during 1968. It
is well modeled by an increase in magnitudes of all events by 0.3 and an increase in the number of events smaller than 4.7 by a factor of 1.6. The cause of this magnitude increase is unknown, but it is presently being studied. Preliminary results suggest that it is related to the closure of the VELA stations which generally had large magnitude biases (North, 1977).

These results strongly suggest that the precursor to the 1971 Valparaiso Chile event recognized by Habermann and reported by Nishenko (1985) is in fact

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a period during which Magnitudes are systematically low relative to the time periods surrounding it and not a real period of low seismicity rate. He are presently in the process of trying to recalculate these magnitudes using station corrections to test this interpretation and to determine whether or not a real precursor preceded this event.

The fact that the detection decrease associated with the closure of the VELA arrays is a global phenomena suggests that the magnitude changes associated with that closure could also be global. Comparison of the magnitude signatures in Figure 2 and 7C suggests that this is so, but that the nature of the change in magnitudes varies regionally. In the Chile case, we know that the change during 1968 involves a decrease in magnitudes from the analysis described above. The small differences between the right hand sides of the magnitude signatures in the middle frame of Figure 7C also indicates that that side of the magnitude signatures reflects primarily magnitude changes in these complex cases. The right side of the magnitude signature for the Kuriles (Figure 2) shows little change, suggesting that the magnitudes in that region did not change strongly during 1968. The right side of the magnitude signature for Central America and Mexico, in contrast, shows rate increases, suggesting an increase in magnitudes at this time in that region. The very high z-values on the left side of that plot probably reflect the combined effect of this magnitude increase and the detection decrease. This is the opposite of the "peak suppression" caused by the magnitude decrease in Chile.

In addition to these cases, observations from the Aleutians suggest that
the magnitudes decreased there during 1968, although a complete analysis has not yet been done. Figure 8 shows the CNET curve for the Central Aleutians from 1963 to 1979. Note the obvious quiet period which occurs very close in time to that described above in Chile. This quiet period in the Aleutians was interpreted by Habermann (1981b) as a precursor to the 1969 Delarof Islands earthquake as well as the 1970 Adak Canyon events. The previous analysis for Chile and the temporal coincidence of these two quiet periods certainly suggest that this interpretation may be in error.

Problems with magnitude stability are not restricted to teleseisaic catalogs. Bakun (1985) presented evidence of systematic changes as large as 0.26 units for events in in the U.S.G.S. CALNET catalog with 2.5 < MD < 3.0. He used magnitudes calculated by Berkeley as a standard and compared the differences between the Berkeley magnitudes and the U.S.G.S. magnitudes. Bakun assumed that the changes in magnitudes reflected changes in gains in U.S.G.S. instruments during April, 1977. Recent analysis of an expanded data set by Craig and Habermann (1986) supports the size of the changes determined by Bakun, but indicates that the picture is much more complicated then previously recognized, with the changes occurring at different times in different regions and magnitude bands.

Recent studies by Habermann have shown that the magnitudes of events smaller than 2.5 are also unstable in the U.S.G.S. CALNET catalog. These results depend on analysis of synthetic magnitude signatures such as those described above. A recent comparison of these results to those from a U.S.G.S. study is described by Habermann (1986) and shown in Figure 9.

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The upper frame of Figure 9 shows the magnitude signature which compares seismicity rates from December 28, 1983 to November 27, 1984 to those between November 28, 1984 and June 4, 1985. This magnitude signature shows the characteristics of a magnitude decrease clearly. The data sets with lower cutoffs, on the right side of the plot, show strong rate decreases. The data sets with upper cutoffs, on the left side of the plot show rate increases.

This magnitude signature was synthesized using the techniques outlined above and the synthetic with the lowest residual, that generated by decreasing the magnitudes of events smaller than 1.9 by .19, is shown by diamonds in Figure 9. The fact that this is the best fit suggests that a decrease of this amount occurred in the magnitudes at this time.

This result could be tested because the magnitudes in this region were recently recalculated using station corrections by Poley (personal communication). The differences between the old and new magnitudes are shown in the bottom of Figure 9 as a function of time. Note the sharp increase in differences (decrease in the old magnitudes) during November, 1984. The change in the old magnitudes can be determined by computing the difference between the means of these two groups of data. This difference is .18 in close agreement with that predicted by the magnitude signature analysis.

Summary

These studies show that systematic changes in magnitudes can cause apparent changes in seismicity rates which can easily be mistaken for precursors. In fact, magnitude instability is a major barrier to using seismicity rates changes for earthquake prediction in both teleseismic and local seismicity catalogs. Despite major advances in understanding of these problems is is still very difficult to resolve them completely in many cases. It seems, therefore, that the only real solution is to redetermine magnitudes for all events in these catalogs using station corrections. This large and difficult task has been undertaken by Bakun and Michaelson (1987) for the CALNET catalog. Work is underway to start the task in the teleseismic catalog. If these studies are successful at removing the systematic biases from the magnitude estimates in these catalogs we will have made a major step toward using the data in earthquake prediction efforts.

FALSE ALARM RATES

Observing some precursor in the region of a mainshock is only the beginning of determining whether that precursor may be useful for earthquake prediction. The second task, given such an observation, is to find other occurrences of similar phenomena and see if they are also followed by mainshocks. If they are, they are successful precursors, if not, they are false alarms. In the case of precursory quiescence, and many other phenomena, it is relatively easy to find the precursors because they are followed by mainshocks. The task of finding the quiescences which are not followed by mainshocks is much more difficult and depends on systematic searches of large amounts of data. A general scheme for approaching this problem has been described by Habermann (1981b) and applied to several cases. These are the

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only systematic studies of false alarm rates for precursory quiescence that I am aware of and I outline the approach here.

One must define anomalies quantitatively in order to search for false alarms associated with those anomalies. The definitions we use involve four parameters which are described in Figure 10. The first two parameters of the definition, the length and the duration, are obvious. The third parameter is a statistic describing the anomaly. In the case of rate changes, the statistic is some measure of the significance of the difference between the two rates. We have used the z-test for a difference between two means, but other statistics can be added to the definition. The first three parameters are all dependent on the magnitude band being examined, hence it must be included as part of the anomaly definition.

The anomaly definitions described above form the basis for an evaluation program. The Coyote Lake case shown in Figure 11 provides a good example of anomaly definitions. All of the data sets shown come from the same segment of the fault which has a length of 21 km. The definition parameters which vary are duration, statistic, and magnitude band. The magnitude bands for each case are given at the top of each frame. In this case the statistic is the z-value which results from the comparison of the background rate to the anomalous rate. The AS(t) functions in this Figure show the significance for all possible durations, therefore they reveal the trade off between duration and significance. These functions indicate that many anomaly durations yield significance levels above 99%; all of those for which the z-value is above the uppermost horizontal line on the scale. All of these durations might yield anomaly definitions with acceptable false alarm rates.

The anomaly evaluation program tests all of these possible anomaly definitions to find the ones with the lowest false alarm rates. This test is done by dividing the fault into a number of segments with the same length as the segment in which the anomaly is defined (see Figure 12). In the usual case we overlap these regions by half of their length in order to increase our coverage (not shown in Figure 12). A window with the duration of the anomaly is slid through the data for each fault segment and compared to all background rates prior to it. The background periods expand to include all of the events prior to the window as it is slid through the data. Any comparison which generates a z-value greater than that of the anomaly being evaluated is an equivalent anomaly.

The equivalent anomalies are then examined and grouped into three groups: Those which occurred prior to past earthquakes are termed precursors to past events; those which are presently occurring are possible precursors to future events, and those that are not associated with past events are false alarms.

This approach to false alarm rate determination has been applied in the Aleutians by Habermann (1981b), in the Tonga-Kermadec by Wyss et al. (1984), and in California by Wyss and Burford (1985) and by Wyss and Habermann (1986). The results of these studies are summarized in Table 1. In general the false alarm searches extend over regions which are at least ten times the size of the anomalous zone and cover time periods ten times the duration of the anomaly. In general the anomalies can be recognized with very few false alarms which is encouraging.

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STATISTICAL APPROACH

The discussion above clearly points out the inadequacies of visual evaluation of seismicity data not only for recognition of anomalies, but for false alarm rate determination. Host workers now agree that some sort of quantitative tools must be used for examining these data. The tools must allow objective comparison of different observed rate changes. It is not clear that an absolute estimate of the significance of a rate change is critical, but relative estimates are important. In addition, the tools must be readily adaptable to efficient computer algorithms, for there is a vast quantity of data which requires analysis. They also must be capable of functioning in a real-time setting with a large data base. Several statistical techniques which are reasonable candidates are outlined below. For complete descriptions of the tests one should examine the original references.

The test which has been most extensively used and tested for comparing seismicity rate changes is the z-test for a difference between two means. This is one of the most general parametric tests for comparing means and the requirements for applicability are very general. In essence, the test is applicable to any large sample of independent data which come from a distribution with a finite mean and a positive variance.

Seismicity rate changes are defined in terms of the rates during the two time periods which proceed and follow the change. In order to compare those rates, we determine the mean rates during the two periods (Ml and H2), the standard deviations of those rates (SI and S2), and the number of samples in each period (Nl and N2). We then calculate the z-statistic for the change using the formula:

(Ml - M2)
2 =

((SI 2/ Nl) *- <S2 2/ N2)> 1/2

Several tests which are based on the assumption that the distribution of seismicity is Poissonian have been used for evaluating the significance of seismicity rate changes. Although it is known that seismicity data are not simply Poissonian (principally because of aftershock sequences), many workers make this assumption in order to allow analysis using Poisson tools. Of course if this assumption is to be made one must somehow remove dependent events from the catalog prior to application of the test (this is also true for the z-test).

If the Poisson assumption is accepted, then one can calculate a mean (M)
for an observed set of data and then determine the probability of observing some number of events (X) during a period with a given duration (N) using the following formula:

e -MN (MN) X / x.

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(Ohtake et al., 1981 and McNally, 1981).

The third test for a difference in rates was proposed by Veneziano and
Van Oyck (1985). This test also includes a Poisson assumption and recasts the test in terms of the binomial distribution. In this case one compares the number of events before and after the suspected change point using the following formula:

nrit ki » / ne \ na , « % ne-na PCNa =na ! Me =ne 3 = ( )p (1-p) na

where Na and He are the expected number of events in the anomalous and entire time periods, na and ne are the observed numbers of events in the two periods, and p is the ratio of the anomalous period duration to the duration of the entire sample.

If the process is stationary, one would expect the ratio of the number of events in the background and anomalous periods to be equal to the ratio of the durations of the periods. The numbers of events in the two time periods are, therefore, distributed as a binomial distribution with parameter p. One observes the number of events that the two time periods do contain and tests for that division using the binomial distribution with that parameter.

A final test which has been used for studying seismicity rates was proposed by Matthews and Reasenberg (1987b). This test is similar in approach to that of Veneziano and Van Dyck (1985), in that it takes advantage of the fact that the number of events in a sub-interval of the data has a binomial distribution with the parameter determined by the duration of the sub-interval. The statistic is given by:

na- Na
s -

(Na (1 - p)) !/ 2

where the symbols are the same as in the last case.

We have applied the first three tests to the same data in a number of cases and found that the resulting estimates of significance of the rate changes are very consistent. The last test has been applied to some of the same data sets by Reasenberg and Matthews (1987) and yields generally lower significance estimates. Understanding the meaning and cause of these differences is an important goal of ongoing work.

Regardless of the test chosen, one must make a choice of what background rate to use when applying the test to data. This choice does affect the resulting significance estimate. Consider the test for occurrence of an observed number of events in a stationary Poisson series. The strictest test is to use the entire period, including the suspected anomaly, as the period over which the background rate is calculated (long background). The other

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choice is to somehow select the suspected anomalous period and use the preceding period as the background estimate. This technique (termed short background), will obviously lead to higher estimated significance of the anomaly because the mean rate of the background process will be higher. This is the approach used by Ohtake et al., 1981 and McNally, 1981 to evaluate the significance of proposed anomalies in Mexico. In this case the anomaly selection is done visually, a difficult technique to apply in large-scale studies of seismicity rate changes.

The problem of objectively identifying the time of anomaly initiation has been addressed in a number of papers by Habermann and others. We developed several windowing schemes which make this possible. The Coyote Lake data in Figure 11 shows the most general scheme for anomaly recognition, the AS(t) function. This function was designed to find the most significant change in rate between two times. It achieves this by testing all possible rate comparisons between those two times. The time of most significant change can then be easily picked from the list. In addition, the AS(t) functions show the trade-off between anomaly duration and significance which is important for the false alarm determination schemes described above.

In addition to testing any given time period for changes, an algorithm
for finding all changes within a data set significant at some level has been developed and described by Habermann (1983). This algorithm makes it possible to analyze a large number of data sets easily. This is clearly important for examining the magnitude distribution of observed changes and for developing the background database necessary for implementing a real-time anomaly recognition and evaluation program.

A different windowing scheme is used in the anomaly evaluation and false alarm recognition part of these studies. This is termed the RAS(t) function (real-time AS). This scheme slides a fixed duration window through the data one sample at a time and compares the rate in that window to the rate in a background window which expands behind the possible anomaly window. This scheme is used because we believe that at any given time one wants to accept the longest possible period for a background estimate.

As stated above, the choice of windowing scheme is independent of the choice of statistic, and there is no reason why the choice should be limited to one of each. The statistics and windowing schemes we have described here are certainly not the only possible choices. The ones we have developed, however, are the most extensively used and tested.

Matthews and Reasenberg (1987a) point out a potential problem which is related to the process of searching for alarms using a statistical approach like those outlined above. The fact that one searches for the anomaly by calculating a large number of statistics decreases the significance associated with a given value of the test statistic. The solution to this problem is to determine the significance of a given value of the test statistic through a series of simulations using data with a known distribution. Habermann and Wyss (1987) discuss this approach in some detail.

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PREDICTIONS AND SUCCESSES

Much of the present optimism associated with changes in seismicity rates
and earthquake prediction comes from recent successful predictions based on quiescence. Of the many possible precursors proposed during the last several decades, quiescence presently appears to be the most promising. This optimism, however, must be tempered by recognition of the fact that successful predictions do not guarantee understanding of the phenomena or event correct analysis of the data. As pointed out above, there are many potential pitfalls involved in using seismicity rates for earthquake predictions and many of the studies which have led to successful predictions have problems which need to be addressed before the success can be considered complete. As pointed out above, the most substantial problems are the elimination of man-made changes as causes of the observations and systematic determination of false alarm rates.

Host seismologists agree that a successful earthquake prediction includes
the time, location, and size of the upcoming event as well as some estimate of the probability of the occurrence of the event. At this point in the development of earthquake prediction, however, our experience is so limited that estimates of the time and probability of occurrence are very poorly constrained. I include as successes, therefore, all cases in which an anomaly is recognized prior to the occurrence of the mainshock. In practice, this is the most important task for a first-pass technique. While we will certainly improve the intermediate-term time estimates which can be made using seismicity patterns by developing more case studies, important information on the time of occurrence will also come from integration of other types of observations with seismicity observations.

Oaxaca, 1978.

The first case of precursory quiescence which was recognized before a mainshock was that prior to the 1978 Oaxaca event (Ohtake et al., 1977a,b, 1981). The authors identified a number of cases in the same tectonic setting and recognized a similar pattern occurring in the Oaxaca region. They predicted the location and size of the event on the basis of the seismic gap between the 1965 and 1968 Oaxaca events. Though the time of the event was not specified, the authors expected a renewal of activity prior to the mainshock. Commentary and analysis of this prediction is included in Garza and Lomnitz, 1978; Haberroann, 1981a; and McNally, 1981.

Mammoth Lake, 1980.

The events during May 1980 in the Mammoth Lakes, California area were predicted on the basis of a period of quiescence in the Sierra Nevada-Great Basin boundary zone by VanWormer and Ryall, 1980. Their prediction was very general: "It is conceivable that this sequence Cof seismicity] could culminate in a major shock with faulting in the zone between Mammoth Lakes and Bridgeport". It is not clear why this region was specified, and the Mammoth Lake events actually were outside of the region (south of Mammoth Lakes rather than north of it), so it is not obvious that this prediction can be considered a success. One point referred to by VanWormer and Ryall (1980) as well as

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Ryall and Ryall (1981) was thai the anomaly was "regional", not restricted to the area of the rupture. I would interpret this as a suggestion that the anomaly was related to network operation rather than to the upcoming earthquake. I believe that this case needs substantial work before the connection between the quiescence and the mainshocks is established.

Costa Rica, 1982.

McNally (1982) described a prediction that was Made for the central region of Costa Rica by Guendel and McNally (1981) as a success. They expected an event with mb ^5.2 in this region between February 1979 and July 1982. An event with MS = 5.6 (unknown Mb) occurred in August 1982 within 50km of the predicted location.

Stone Canyon, 1986.

The location and size of an event with ML = 4.9 which occurred on May 31,
1986 near Stone Canyon, California was successfully predicted by Wyss and Burford (1985). This is perhaps the Most complete prediction Made to date on the basis of quiescence. The seismicity of the segment of the San Andreas fault near Stone Canyon as well as the catalog available for the region are complex and difficult to work with. Wyss and Burford faced a number of problems which had not been addressed in most previous studies. These included man-made changes in the catalog, false alarm recognition, and complex spatial relationships between several quiet fault segments. This work is discussed in a number of companion comments published with Wyss and Burford' paper as well as by Wyss and Burford, 1986.

Andreanof Islands, 1986.

The May 7, 1986 event in the Andreanof Islands (MS =7.7) was predicted
by Kisslinger, 1985. Much of the rupture zone of this event was in the area monitored by a local network centered on Adak Island in the Aleutians. The local seismicity in the area of the network showed a sharp decrease in rate during September, 1982. This quiescence prompted a number of studies in the region related to the possible occurrence of a large event. This case combines aspects of local and teleseismic data and, therefore, will provide many important data in the near future. The prediction is reviewed by Kisslinger, 1987.

Acapulco

McNally, 1981 presented evidence which indicated that the area near Acapulco, Mexico became quiet during January 1977. While she concluded that "our current knowledge is not sufficient to predict to predict the exact time, magnitude, and location of future earthquakes; a special study of the area is warranted." Thus no actual prediction was made but there were strong implications that an event was expected. This observation is discussed above.

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Hellenic Arc

Wyss and Baer, 1981 examined the seismic potential of the Hellenic Arc south of Greece and concluded that the entire arc had potential for a large or great earthquake based on the period since the last such events. In addition, they concluded that several sections of the arc had been quiet since 1962 and 1966. They proposed that these sections of the arc would experience large to great earthquakes before 1991.

San Juan Bautista, California

Wyss and Burford (1985) recognized a segment of the San Andreas fault in
the vicinity of San Juan Bautista (near 36.84 N which had been quiet since February 1983. They proposed that this segment of the fault would rupture soon in a ML = 5.0 + 0.5 event. The analysis in this case was similar to that in the Stone Canyon 1986 case but the data need to be reevaluated as the expected time of occurrence of the event has been passed.

Bear Valley, California

Wyss and Habermann (1986) examined seismicity rates on the segment of the
San Andreas fault between 36.3 and 37N and concluded that a 10km long segment of the fault near 36.4 N had been quiet between Nay, 1985 and June, 1986 (the end of their data). This study included an extensive examination of man-made changes in the catalog as well as false alarm rates. The possible precursor they recognized is very similar to that described by Wyss and Burford (1985) before the Nay 31, 1986 event. They conclude that this segment of the fault will rupture in an event with ML = 5.0 + .5 between December 1986 and January 1988.

Summary

The successes and predictions outlined here provide a broad foundation
for further work as well as a wide spectrum of quality and completeness. Nost have a number of rather obvious shortcomings which include lack of quantitative evaluations of the anomalies, failure to correct for man-made changes in the catalog, and lack of a systematic search for false alarms. These shortcomings make it difficult to realistically evaluate most of the results at this point and indicate important avenues for further work. Nevertheless, the successes indicate that real anomalies are often resilient to errors in analysis and that there is reason to be optimistic about future
success.

FUTURE DIRECTIONS

Analysis of Existing Data

As pointed out in the section on false alarm recognition, it is

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relatively easy to study seismicity rates prior to past mainshocks and to determine whether precursory quiescence occurred. This, however, is only half of the task. The second aspect of the task involves finding the quiet periods which are not associated with mainshocks. This requires systematic analysis of a large amount of seismicity data. These data exist and are presently underanalyzed because of lack of resources.

Network data for all of California, Alasks, and Hawaii is collected directly by internal investigators at the the U.S.G.S. External investigators collect data in a wide variety of locations including Adak, the Shumagins, Utah, Missouri, the New Hebrides, Washington, and the Caribbean. All of these data must be systematically analyzed if we are to understand the processes responsible for rate changes and the predictive capability of these changes. The analysis of these data should include:

* Analysis of the detection and reporting histories of the networks to identify and correct for man-made changes in the catalog. This task
may include redetermination of magnitudes if it is found that the present magnitudes are not stable over the duration of the catalog.

* Examination of seismicity rates prior to all past mainshocks in the catalogs (M > 5.0) to determine whether they were preceded by seismic quiescence and to develop a series of anomaly definitions.

* Systematic false alarm recognition studies using the anomaly definitions developed during the last step. This task would also include the development of a seismicity rate database which could be searched systematically in the event of a new anomaly definition.

* Development and implementation of real-time anomaly recognition software capable of examining seismicity data as it is collected and checking for occurrence of a number of anomalies (as defined above)
using a variety of statistical tools.

It is clear that this analysis can not be accomplished overnight and that
it will involve a great deal of work, but the database generated by the study will be invaluable. It will allow us to determine empirically on the basis of a large temporal and spatial database what the false alarm rate associated with any given observation is. Of course this analysis should not be limited to study of seismic quiescence, but should include examination for other possible seismicity precursors (clusters, foreshocks, migration, activation, etc.). It is important that a number of statistical approaches be included in the study as well, to insure balance and objectivity.

Standardization of Analysis and Evaluation

Several predictions based on quiescence were presented to the National Earthquake Prediction Evaluation Council recently and in both cases the panel was uncertain about how to evaluate the predictions and concluded that the phenomenon required more study. Predictions which are based on quiescence are likely to become more complex in the future and this complexity will add to the difficulty of evaluating these predictions. In order to alleviate this

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problem we need to agree on standards which predictions based on quiescence must adhere to. Possible standards are:

* The prediction must demonstrate that the observed precursor is not man-made. This could be possibly be done by showing that the precursor
has a limited spatial extent and that nearby regions have no quiescence
or it could be done using a magnitude signature analysis similar to
those outlined above.

The distribution of the change in the magnitude domain must be specified. This analysis should include examination of a number of magnitude bands with both upper and lower cutoffs.

The prediction must include a demonstration of magnitude and detection stability in the data set used to recognize the precursor and in the false alarm test.

* A systematic search for false alarms in a region ten times the size of the anomaly must be included in the prediction.

The study should show that at least three statistical evaluations of the anomaly indicate consistently high significance.

* Important supporting evidence from tectonics, repeat time information
or seismic gaps, other seismological studies (stress drops, focal mechanisms, velocities), geodetics, etc. should be provided (if possible),

In addition to standardizing the requirements for a prediction, we must
aim toward standardizing the techniques of evaluating predictions. At present the evaluations result from generally informal discussions of the National Earthquake Prediction Evaluation Council. While these discussions are certainly worthwhile, it is difficult to systematically evaluate and thereby improve the evaluations. Such second level evaluation is clearly necessary and will become even more so as the complexity of the predictions increases. A number of techniques of analysis of expert opinions have recently been incorporated into studies of seismic risk in situations where "soft" data must be evaluated in relation to problems with many unknowns. It will be relatively easy to apply the same techniques to the evaluation of earthquake predictions. This should be done as an adjunct to the present system as soon as possible.

Tectonic Information and Fault Properties

We need to increase our understanding of the tectonic processes and fault properties in the regions where we are attempting predictions. In some cases, such as in California, this task can be carried out in some detail because the faults are easily accessible. In subduction zone settings we must improve our capability to infer fault properties from seismicity distributions and source parameters of smaller events. Such work was important in the recent prediction of the Andreanof Island event, although the focus was on the wrong section of the plate boundary.

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Fault geometry may be very important in controlling the distribution of stress and strength on faults. In strike-slip situations fault we should take advantage of our knowledge of fault geometry to develop specific models of seismicity precursors and test those models. In the event of the observation of a well documented case of seismic quiescence on a segment of a fault we should be able to mobilize a field study quickly in order to gain crucial information on the almost entirely unknown mechanism of seismic quiescence.

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Table 1. Results of False Alarm Searches
Anomaly Number of Number of
Hainshock Length Duration z-value Comparisons False Alarms

Delarof Is. 1969
Adak Canyon 1971
Delarof Is. 1975
Delarof Is. 1978
Tonga 1977
Stone Canyon 1982
Stone Canyon 1986
San Juan Bautista
Cienega Winery

(km)
135
67
67
67
190
10
7
5
10

(mo)
21
33
15
48
25
19
65
99
130

3.5
3.0
2.3
2.7
4.5
3.2-4.0
6.1-7.7
2.8-3.3
2.8-3.3

^2000
3400
4248
3840
3360

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Figure 1. Cumulative number of events reported in the NOAA Hypocenter Data File for three magnitude bands. Tines of changes in rate are indicated by vertical bars. ALL events are all events with »b's assigned. Note the rate decrease in this set during 1968 (dashed line shows the 1964 to 1968 rate). Because the entire world is considered, this decrease must be related to a decrease in detection. Events with mb ^ 4.5 show a constant rate of listing since 1963. This is interpreted as an indication of constant occurrence, detection, and reporting rates for these relatively large events. Events with Mb <_ 4.4, in contrast, show significant rate decreases during 1968 and 1976. These decreases, especially the one near 1968 emphasized by the dashed line, control the shape of the curve for ALL events. From Haberaann, 1982a.

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Figure 2A. Magnitude versus time diagram of shallow earthquakes which took place in the Oaxaca quiescent zone between 1963 and 1978. Each event reported by the PDE is plotted by a vertical segment with length proportional to the body wave magnitude. The periods of quiescence (beta) and activation (alpha) are marked. Note the strong decrease in the number of small events reported in this region during 1968. The levels of detection between 1963 to 1968 and 1968 to 1973 are certainly different. From Ohtake et al. f 1981.

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Figure 3. Space-time plot of the seisaicity of the Kurile Is. region. See Figure 2B for details. Note the vertical stripe of high activity in this region during 1961. This reflects inclusion of information from a Moscow catalog for that year. From Kanamori, 1981.

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Figure 4. Cumulative nu.ber of events («b > 4.0) versus tl.e for the Acapulco region. Arrows Indicate .ainshocks. Rates nor.allzed by area are shown and Irfbased on the tl.e periods for which the rates are shown as "Id lines MM. the effect of the detection decrease In this region during 1968 Undlcatd by the h^avy dashed line added to the Figure). Fro. McNally, 1981.

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Figure 5. Yearly number of seismic events (M >6) reported to have occurred in the focal and adjacent regions of the 1952 Kamchatka, 1957 Aleutians, and 1964 Alaska great earthquakes. Earthquake data are taken at face value from the ISC. Note the large number of events reported in 1930-1935 (Gutenberg and Richter's special effort to find earthquakes with M > 6). Also note the increased number of events when BCIS and PDE started reporting, and after the efforts made during the International Geophysical Year (IGY) in 1957-58. Note also the lack of events with M < 7 prior to 1930 and after 1935 (World War II). All these increases and decreases in seismicity (M < 7) are interpreted to be due to variations in the world's detection and reporting capabilities (Perez and Scholz, 1984). Reporting for larger earthquakes (M > 7) appears to be less affected by these variations. From Perez, 1983.

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Figure 6. Cumulative number of all large earthquakes in the world during this century for three different magnitude bands. All three bands clearly show strong changes in rate (slope of the curves). Note that the rates between the obvious slope changes are very linear, suggesting that events are occurring at constant rates when averaged over the whole globe. The changes in slope, must, therefore, reflect changes in the detection or reporting of these events. These apparent changes in rates can be explained if the magnitudes prior to 1908 were overestimated by at least 0.5 units and those between 1908 and 1948 were overestimated by 0.2 units. From Perez and Scholz, 1984.

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Figure 7B. Cumulative number of events in a 1900 km long segment of the Chilean subduction zone which includes the regions covered in Figure 7A. The magnitude cutoff in this case is 4.8. Note the clear apparently quiet period between March 1968 and July 1970. The fact that this quiet period is clear in such a large segment of the seismic zone suggests that it is due to a change in the detection or reporting system. The fact that it affects larger events strongly eliminates the possibility of a detection decrease. It Bust, therefore, be a systematic decrease in magnitudes.

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Figure 7C. Magnitude signature comparing the rates between July 1964 and March 1968 to those between April 1968 and July 1970 in the Chilean seismic zone between 22 and 39S. The observed magnitude signature is shown by boxes in all three frames. The asterisks in the middle frame show the magnitude signature which results from decreasing the magnitudes of all events by .4 and comparing the modified background to the original. This shows the expected decreases in magnitude bands with lower cutoffs (on the right side of the plot) and increases in magnitude bands with upper cutoffs (on the left). Note that the form of this synthetic is very similar to that of the observed (on the right side of the plot), but the observed magnitude signature indicates rate decreases in the data sets with upper cutoffs. This is simulated by decreasing the number of events smaller than 4.6 by a factor of two. The resulting magnitude signature is shown by diamonds in the middle frame. This fits most of the observations quite well except for the smallest events. The misfit indicates that the magnitude decrease was too strong for the small events, so next the magnitudes of the events smaller than 4.3 were increased by 0.3. The final synthetic which combines all three of these changes is shown by the diamonds in the bottom frame. We conclude that the change during March 1968 was a combination of a detection decrease and a decrease in magnitudes which was stronger for the larger events.

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Figure 7D. Magnitude signature comparing the rates between August 1969 and July 1970 to those between August 1970 and October 1971 in the Chilean seismic zone between 22 and 39S. The observed magnitude signature is shown by boxes in all three frames. It is a rather simple change which has the characteristics expected for an increase in magnitudes: rate increases in magnitude bands with lower cutoffs (on the right side of the plot) and rate decreases in magnitude bands with upper cutoffs (on the left). The synthetic magnitude signature generated by increasing the magnitudes of all events by .3 is shown by diamonds in the middle frame. This change clearly accounts for most of the observations. The slightly negative platform on the left side of the plot, however, indicates a small increase in detection, This is simulated by increasing the number of events smaller than 4.7 by a factor of 1.6. The final synthetic magnitude signature which combines these two effects is shown by diamonds in the bottom frame.

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TIME, YEARS

Figure 8. Cunulaiive number of events as Aleutians between 1963 and 1979. Note the 1970 in the data for events larger than 4.
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Figure 9. The upper frame of Figure 9 shows the magnitude signature which compares seismicity rates from December 28, 1983 to November 27, 1984 to those between November 28, 1984 and June 4, 1985 in the pre-1985 Lindh catalog for the Parkfield region of California (boxes). This magnitude signature shows the characteristics of a magnitude decrease clearly. The data sets with lower cutoffs, on the right side of the plot, show strong rate decreases. The data sets with upper cutoffs, on the left side of the plot show rate increases. The diamonds show the synthetic magnitude signature generated by decreasing the magnitudes of all events smaller than 1.9 by 0.19. This change in magnitudes clearly explains the observations quite well. The bottom frame shows differences between magnitudes calculated for these data using station corrections and the original magnitudes as a function of time. Note the clear systematic change in these differences during November 1984, the time of the change examined in the magnitude signature. The change in mean differences is . 18, in close agreement with the change found using the synthetic magnitude signatures.

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Anomaly Definitions Anomalies must be quantitatively defined in order to be evaluated. Our definitions

include four parameters:

1 . Length
The length of an anomaly is the length of the fault segment which experiences the anomaly.

LENGTH
Fault

2. Duration

The Duration is the time from the beginning of the anomaly to the mainshock This part of the definition can also include the duration of the background period used to recognize the anomaly.

3 Statistic
The difference between
the background and
anomalous rates is quan-
tified using the z-test for
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4. Magnitude Band
All of the other parameters depend on the magnitude band which is being used for the anomaly evaluation, so it is an important part of the definition.

Figure 10. Parameters included in anonaly definitions,

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Figure 11. Cumulative nunber of events as a function of ti«e for a segment of the Calaveras fault including the rupture zone of the Coyote Lake earthquake of July 6, 1979. The aainshock occurs at the end of the curves. Also shown are AS(t) functions which give the z-values resulting from the comparisons of the rates between the beginning of the data set and t to those between t * 1 and the end of the quiet period prior to the nainshock. Note that significant quiet periods exist prior to the aainshock for all three magnitude cutoffs.

Anomaly Evaluation

Identifying an anomaly is the first step, Evaluating the anomaly is the process of finding all occurrences of the anomaly, including those not followed by mainshocks (false alarms).

The anomaly is observed and defined in the
\%f%Mffiffl(segment of the fault. This segment
has the same length as the anomaly, and a
period of anomalous seismicity rate with a
known duration has occurred there. The significance of this anomaly is determined by some statistic, and the magnitude band in which the anomaly occurred is known. Thus, the anomaly is completely specified.

The Y//////A segments of the fault are all possible sites of anomalies of the same length, duration, statistic, and magnitude band which we term equivalent anomalies. Anomaly evaluation is the process of searching the segments for these equivalent
anomalies. When they are found they are classed as precursors to past events, possible precursors to future events, or false alarms.

Figure 12. Anomaly evaluation.

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DOWNWARD MIGRATION OF SEISMIC ACTIVITY PRIOR TO SOME GREAT SHALLOW EARTHQUAKES IN THE SUBDUCTION ZONES IN JAPAN - A POSSIBLE INTERMEDIATE-TERM PRECURSORY PHENOMENON

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Abstract

Before the 1944 Tonankai earthquake along the Nankai Trough, seismic activity increased in the shallow depths, and then the activity gradually migrated downwards. When it reached its limit (a depth of approximately 70 km), the main shock occurred. Several deep earthquakes, including one of M 5.3, occurred several months prior to the Tonankai earthquake. A similar downward migration pattern can be also recognized in the case of the 1952 Tokachi-oki earthquake. In this case the deepest earthquakes reached a depth of about 400 km. This may be one of the intermediate-term precursory phenomena of great thrust-type earthquakes in subduction zones. Recent observations in the Tokai district along the Suruga Trough where a large earthquake is expected to occur in the future suggest a similar downward migration pattern in the land area.

Introduction

Various long-term precursory phenomena, such as a gradual buildup in seismic activity over a wide area and the appearance of a seismic gap of the second kind (quiescence in the focal region) and the doughnut pattern (increased activity surrounding the focal region), are often observed prior to a large earthquake (e.g. Mogi, 1985). It has been pointed out that the relation log T = aM + b exists between the duration (T) of long-term precursory phenomena and the magnitude (M) of the subsequent earthquake (Tsubokawa, 1969; Whitcomb et al., 1973; Scholz et al., 1973; Rikitake, 1975, 1979; Ohtake, 1980; Mogi, 1986d), so if the value of M can be estimated it will be possible to infer the time of occurrence of the earthquake. In actual practice, however, the value of T varies greatly, so inevitably this estimate is a very rough one.
There are also many cases in which various kinds of precursory
phenomena have been observed immediately before an earthquake, such as foreshocks and anomalous crustal movements (e.g. Rikitake, 1979). Most of these short-term precursors appear several days or several hours before the earthquake, and observing them may make it possible to predict an earthquake just before it occurs. It is well known that there have been successful cases of earthquake prediction using foreshocks as the deciding factor, such as the Haicheng earthquake in China (e.g. Raleigh et al., 1977).
It is hoped that a method for intermediate-term earthquake
prediction, linking long-term forecasts and short-term prediction immediately prior to earthquakes, will be developed. This means

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developing a method for predicting in the order of years or months the process by which an earthquake gradually becomes imminent.
This paper discusses the downward migration of seismic activity along
the deep seismic plane prior to large shallow earthquakes along subduction zones, which may be a lead to intermediate-term earthquake prediction. We will describe how this phenomenon has been recognized in the case of the 1944 Tonankai earthquake and the 1952 Tokachi-oki earthquake. If this phenomenon is recognized for other large earthquakes along subduction zones, it may act as an effective method of intermediate-term earthquake prediction.

1944 Tonankai earthquake

Large M 8 class earthquakes have occurred repeatedly along the Nankai Trough at regular intervals of about 100-150 years. In the 1707 Hoei earthquake a rupture occurred in the whole region from the Suruga Trough to the Nankai Trough. One hundred and fifty years later on December 23, 1854 the Ansei Tokai earthquake occurred in the eastern half of this region, followed the next day (December 24) by the Ansei Nankaido earthquake in the western half of the region. This century has witnessed the 1944 Tonankai earthquake (M 7.9) in the eastern half of the Nankai Trough region and the Nankaido earthquake (M 8.0) in the western half in 1946. The focal region of the 1944 Tonankai earthquake did not reach as far as the Suruga Trough to the east, and this area remains unruptured (seismic gap of the first kind), so that a "Tokai earthquake" is expected to occur there in the near future (e.g. Mogi, 1986a).
Figure 1 shows how the Philippine Sea Plate subducts under the
Eurasian Plate. This subduction commences from the Nankai Trough - Suruga Trough and the Sagami Trough. The subduction zone along the Nankai Trough
- Suruga Trough is divided into three blocks - Nankaido, Tonankai and Tokai - and the 1944 Tonankai earthquake and 1946 Nankaidc earthquake occurred in the Tonankai and Nankaido blocks, respectively.

Changes in long-term seismic activity

In earlier papers we focused on shallow earthquakes of M 6.0 or
larger, and reported that from about 20 years before the 1944 Tonankai earthquake and the 1946 Nankaido earthquake their focal regions became quiescent (appearance of a seismic gap of the second kind).
Simultaneously the surrounding areas, such as the Japan Sea coastal area, which had been quiescent until then, became seismically active (appearance of the doughnut pattern)(Mogi, 1969; 1981). This time we have examined earthquakes of M 5.0 or larger using the latest earthquake catalog issued by the Japan Meteorological Agency (JMA), and reached a similar conclusion, The upper diagram in Fig. 2 shows the distribution of earthquakes during the two decades before the large earthquakes. The lower diagram shows the distribution of the main shocks (ringed solid circles) and the aftershocks of the Tonankai and Nankaido earthquakes. It is noteworthy that the focal regions (indicated roughly by the aftershock regions) of these two great earthquakes were quiescent during the preceding two decades, whereas striking activity occurred in the surrounding areas. However, the lively activity in the region from the western Kii Peninsula down to eastern Shikoku catches the attention. As shown in Fig. 1, this region is located

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at the boundary between the Tonankai and Nankaido blocks, and this activity is regarded as precursory activity to the large earthquakes in 1944 and 1946 (Mogi, 1981). We will discuss the details of this activity in the following section.

Downward migration of seismic activity from the shallow depths

As Fig. 1 shows, the Philippine Sea Plate subducts below the Eurasian Plate along the Pacific Coast of Honshu, and the resultant active seismic plane reaches a depth of 60-70 km. This land area where slightly deep earthquakes occur is located to the north of the seismic gap that existed before the 1944 Tonankai earthquake, and an examination of the earthquakes occurring there reveals that their depths seems to have changed systematically prior to the large earthquake (Mogi, 1986c).
We divided this region into three districts shown in Fig. 3 and
surveyed them. The triangles mark the location of the JMA seismometers used to determine the hypocenter, and they cover the region surveyed quite throughly. Hence it is thought that the determination of the hypocenters is relatively accurate, and that it is possible to discuss their depth (Mogi. 1986c).
Figure 4 shows M-T graphs (M denoted on the vertical axis, time on the horizontal axis) for each depth level for the westernmost of these three districts - i.e. the area from the Kii Peninsula to eastern Shikoku (the Nankaido-Tonankai region in the figure). The uppermost graph is for shallow earthquakes with a depth of 0-30 km. Activity gradually increased from about 1926, peaking with an earthquake of M 6.4 in 1936 and one of
M 6.8 in 1938, and then activity declined slightly. The seismic activity between 1935 and 1940, including these two earthquakes, is particularly marked for this region, and is worthy of attention.
The middle graph is for earthquakes that occurred at a depth of 40-50
km. Activity increased from about 1935 onward, peaking between 1940 and 1942 with two earthquakes of M 5.7 before declining. The bottom graph is for earthquakes at a depth of 60 km or more. In the years leading up to the major earthquake in 1944 activity gradually increased, with a large earthquake of M 6.3 occurring in 1940. During the months prior to the Tonankai earthquake five earthquakes of M 3.5 or larger occurred, the largest of which was M 5.3. This striking succession of earthquakes in the deepest part of the Kii Peninsula was unprecedented. These results show that seismic activity gradually increased prior to the Tonankai earthquake, but that the deeper the hypocenter, the later did the activity peak, with the activity in the deepest part increasing several months before the large earthquake.
In Fig. 5 the same data are used to plot the depth of the hypocenters
on the vertical axis and time on the horizontal axis. The size of the circles indicates the magnitude of the earthquake. There is considerable scattering, but this figure shows that on the whole seismic activity gradually migrated downwards, leading up to the M 7.9 Tonankai earthquake. The Eastern Tokai region shown in Fig. 3 is located adjacent to the
focal region of the Tonankai earthquake. Figure 6 shows the temporal variations in the depth of the hypocenters for this region, as in Fig. 5. In this case there was considerable activity at shallow depths in about 1935, but this activity gradually migrated downwards as time passed. It is very interesting that this downward migration pattern occurred at

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virtually the same time as that in the Nankaido-Tonankai region, suggesting that this change was related to the Tonankai earthquake. In this case, though, few earthquakes occurred at depths below 40 km, and the migration pattern disappears. This may be because the rupture zone of the Tonankai earthquake did not rench as far as the eastern part of the Tokai region, so the activity did not extend down to the deepest part. Since this downward migration of seismic activity did not occur afterwards in normal times (Mogi, 1986c), it can be regarded as a precursor of the Tonankai earthquake.

Summary of precursory phenomena

Sato (1977) has pointed out that anomalous ground tilt occurred at Kakegawa, which is located at the northeastern tip of the focal region of the Tonankai earthquake, and that this was a precursory change occurring immediately before the earthquake. Mogi (1984/85) has reported that these ground disturbances began two or three days before the earthquake, and gradually accelerated (Fig. 7).
Figure 8 collates the time of appearance and duration of the changes regarded as precursors of the Tonankai earthquake, with the horizontal axis denoting time. The top line shows the seismic gap of the second kind and the doughnut pattern, which appeared about 20 years before the Tonankai earthquake and continued up until its occurrence. The next three lines show precursory seismic activity in the land area mentioned earlier, such as the Kii Peninsula. The deeper the hypocenter, the later does the activity peak. The solid circles on this line indicate particularly large earthquakes, and the open circle indicates an earthquake sequence. The next line shows activity along the trough - i.e. where plate subduction commences. Activity here increased five or six years before the large earthquake (Mogi, 1986c). The following line indicates the earthquake swarm near Owase on the southeast coast of the Kii Peninsula as reported by the Association for the Development of Earthquake Prediction (1983). However, the original records have been lost, so there are many unclear points about the swarm. The last line shows the ground tilting at Kakegawa immediately prior to the Tonankai earthquake.
Tonankai and Nankaido earthquakes have occurred repeatedly in the
past at roughly the same locations and with about the same magnitude. When such regularity exists, it is highly likely that the precursory phenomena will also recur in a similar manner. Hence the results collated in Fig. 8 will be of great value in forecasting the occurrence of the next Tonankai earthquake. The pattern whereby seismic activity migrates downward and a large earthquake occurs when this has reached the maximum depth may act as an important lead in intermediate-term earthquake forecasting.

1952 Tokachi-oki earthquake

As described above, seismic activity migrated downwards prior to the 1944 Tonankai earthquake. The next question is whether or not this pattern can be recognized in the case of other large earthquakes also. Previously the author has reported several instances in which activity at the maximum depth of the deep seismic plane increased one or two years before large shallow earthquakes that occur along the Japan-Kurile Trench

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(Mogi, 1973). Among these cases, seismic activity along the deep seismic plane is continuous from the shallow depths to the maximum depth in the region of the 1952 Tokachi-oki earthquake (M 8.2). Accordingly the author examined the temporal variations in the depths of the hypocenters of earthquakes along the deep seismic plane for this earthquake. There have also been reports that an obvious seismic gap of the second kind appeared at a shallow depth before this large earthquake (Inouye, 1965; Ustu, 1968; Mogi, 1969; Katsumata and Yoshida, 1980).
The diagram on the righthand side of Fig. 9 plots the earthquakes in
this region, with the vertical axis denoting the focal depth and the horizontal axis denoting time. The earthquakes plotted in this diagram are those within the range shown in the lefthand diagram of this figure. An examination of this figure makes it clear that deep seismic activity in this region was very active one or two years before the Tokachi-oki earthquake, as Mogi (1973) has already indicated. Shallow seismic activity was marked about ten years before the Tokachi-oki earthquake, but this showed a tendency to migrate downwards as time passed. Deep seismic activity commenced subsequent to this downward migration, including a large earthquake in 1950 (M 7.8, depth of 320 km). This process, whereby seismic activity migrates downward and a relatively large earthquake occurs when it reaches the maximum depth, followed shortly afterwards by a large shallow earthquake, resembles that in the case of the Tonankai earthquake. With the Tonankai earthquake, however, the maximum depth was 60-70 km, whereas it reached down to 400 km in the Tokachi-oki earthquake. Mogi (1973) has discussed the mechanism by which deep seismic
activity increases prior to a major earthquake along a trench and how the large earthquake occurs at the maximum depth, followed shortly by a large shallow earthquake.

Recent seismic activity in the Tokai district

As described in section 2, the Tokai district along the Suruga Trough
has not ruptured in recent years, so it is a likely site for a large earthquake in the near future (e.g. Mogi, 1985). Up until now, however, no changes that are regarded as precursors of such a large earthquake have been reported. Recently, however, the following changes have occurred in the seismic activity in this area (Mogi, 1986b).
(1) Recently several large earthquakes of M 6.5-7.0 have occurred in the area surrounding the focal region of the anticipated "Tokai earthquake", The upper righthand side of Fig. 10 shows an M-T graph for the shallow earthquakes that have occurred in region A in the diagram on the
lefthand side of Fig. 10. Region A, which corresponds to the area surrounding the expected focal region, had been relatively quiescent since 1950, but a series of large earthquakes have occurred since about 1970. These include the 1974 Izu-Hanto-oki earthquake (M 6.9), the
1978 Izu-Oshima-kinkai earthquake (M 7.0), the 1980 Izu-Hanto-toho-oki earthquake (M 6.7) and the 1983 Nagano-ken-seibu earthquake (M 6.8).
(2) The area along the Suruga Trough - Nankai Trough, which includes the expected focal region of the coming "Tokai earthquake", has become quiescent in recent years. The lower right portion of Fig. 10 shows an M-T graph for shallow earthquakes that have occurred in region B in the lefthand diagram. Virtually steady activity had continued since 1950 (despite a break in about 1960), but since 1973 almost no earthquakes

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of M 4.2 or larger have occurred. The 1944 Tonankai earthquake has already occurred in the western half of region B shown here, and our interest at present lies in the eastern part containing Suruga Bay. The reason for including the whole area of region B is that the activity within this region varies in almost the same manner.
Figure lla and b shows the distribution of shallow earthquakes of
M 4.2 (a value determined from Fig. 10) or larger during the two periods 1961-1972 and 1973-1985. Focusing on the area (region B) along the trough marked by the curved line, considerable activity occurred between 1961 and 1972, but the period 1973-1985 has been markedly quiescent. It is noteworthy that this quiescence in region B and the increased seismic activity in region A appeared at virtually the same time, because it may indicate the appearance of a seismic gap of the second kind and the doughnut pattern, which are long-term precursors of a large earthquake.
(3) Somewhat systematic changes can be seen in the depth of the hypocenters of the small to medium earthquakes that have occurred in the land area
of the Tokai district. Figure 12 shows the temporal variations in the focal depth of the earthquakes that have occurred within the block
marked "Tokai" in the upper diagram. There is a trend for the depth of the earthquakes to increase gradually since about 1975. Such systematic changes in depth have not been observed since about 1953, when the after-effects of the 1944 Tonankai earthquake more or less ceased. This tendency is noteworthy because it resembles the above changes observed prior to the Tonankai and Tokachi-oki earthquakes.
Thus several changes in recent seismic activity in the Tokai district
have appeared virtually simultaneously, so it is conceivable that they may be precursory changes of a large earthquake. It is also possible, however, that these changes may merely have appeared as a result of fluctuations in seismic activity, quite unrelated to the occurrence of a large earthquake. At all events, future developments should be watched with care.

Conclusion

Seismic activity migrated downwards prior to the 1944 Tonankai
earthquake (M 7.9) and the 1952 Tokachi-oki earthquake (M 8.2), which
were large shallow earthquakes that occurred in a subduction zone. A somewhat large earthquake occurred at the maximum depth and was followed shortly afterwards by the large shallow earthquake. This downward migration pattern is not always very obvious, but it is noteworthy that it is common to both of these earthquakes. If this pattern can be recognized in other earthquakes also, it may act as a means of forecasting the time of occurrence of large earthquakes. Such intermediate-term prediction would fill the gap between long-term forecasting based on such phenomena as seismic gap of the second kind and earthquake prediction immediately before earthquakes, based on such changes as crustal movements and foreshocks.
Looking at temporal changes in the depth of earthquakes in the land
area of the Tokai district, it is noteworthy that a downward migration pattern of seismic activity has occurred in the past decade. It is possible that such a pattern may have appeared by chance as a fluctuation in seismic activity, so future developments should be carefully watched.

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Fig. 1 Configuration of the deep seismic plane accompanying the subduetion of the northern tip of the Philippine Sea plate, compiled from Yamazaki and Ooida (1985) and Mizoue et al. (1983). The boundaries between the Nankaido, the Tonankai, and the Tokai blocks along the Nankai-Suruga Trough are indicated.

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Fig. 2 Epicentral distribution of shallow earthquakes (focal depth£ depth 60 km) in western Japan. Above: January 1, 1927~ December
6, 1944 (just before the Tonankai earthquake). Below: December 7, 1944~ December 31, 1947. Double circles show the main shocks of
the Tonankai and Nankaido earthquakes. Data are from JMA catalogue,

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136' 138' 140° E

Fig. 3 Areas along the Nankai-Suruga Trough of which seismic activities were surveyed. Triangles: JMA's seismic stations.

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Fig. 5 Focal depths of earthquakes in the Nankaido-Tonankai block are plotted against time. Hypocenters of earthquakes gradually deepened before the Tonankai earthquake as time passed.

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Fig. 6 Focal depths of earthquakes in the eastern Tokai block are plotted against time.

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Tononkai Earthquake

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Fig. 7 Precursory crustal movement (ground tilting) at Kakegawa at the northeastern tip of the focal region of the Tonankai earthquake which was obtained from leveling data (Mogi, 1984/85).

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Time (Years) -

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Seismic gap
Doughnut pattern

Seismic activity
in the land area

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h>60km

Seismic activity along the Mankai Trough

Earthquake swarms at Owase) (100-120 days

Ground tilting at Kakegawa (2-3days)

1920 1925 1930 1935 1940 1945

Fig. 8 The time of appearance and the duration of phenomena regarded as precursors of the Tonankai earthquake.

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Fig. 9 Focal depths of earthquakes before and after the 1952 Tokachi-oki earthquake of M 8.2 in the area shown in the left figure are plotted against time.

PRECURSORY SEISMIC QUIESCENCE

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Abstract

Seventeen reliable cases of precursory seismic quiescence to mainshocks with magni-tudes from Mr = 4.7 to M^.^8.0 have been reported in the literature. The amount of rate decrease ranges from 45% to 90%. The significance of these changes are mostly very high, with the standard deviate z-value exceeding 3.2. The assumption that the back-ground rate is approximately constant is fulfilled in most crustal volumes studied. All quiescence anomalies seem to have abrupt beginnings, and the rate during the anomalous time is fairly constant. The duration of the precursors range from 15 to 75 months, but it is not clear what factors determine that time. At least three successful predictions have been based on seismic quiescence. These cases have shown that mainshocks can be predicted based on quiescence, but they have also shown that the interpretation of the data in real time is difficult and non-unique. The false alarm rate appears to be low, probably less than 50%. Failure to predict may be expected in perhaps 50% of mainshocks even in carefully monitored areas. Quiescence cannot be used as a precursor in tectonic environments with low microseismic activity. Most characteristics of the phenomenon are still poorly defined, but data exist which probably permit at least a doubling of the presently available data on case histories.

Introduction

Seismic quiescence may be the most promising intermediate term precursor. For this reason one needs to assess our present understanding of this phenomenon and to develop a data base of high quality case histories. Far reaching conclusions cannot be drawn from the small number of high quality examples available. Some quiescences reported in the literature allow doubts as to whether they are natural phenomena related to tectonic processes, or whether they are artificial or random rate fluctuations. Cases for which any doubts may exist, in our judgment, were not included in this summary. As we are in the process of forming an opinion on the nature of seismic quiescence, we think it better to form this opinion on the basis of a small but high quality data set, rather than on one diluted by possibly invalid data points. As a consequence we may have left out some data which are in fact valid, and just appear to be lacking in our subjective judgment. Perhaps such data points could be added in the future.

Our definition of quiescence follows: seismic quiescence is a decrease of mean seismi-city rate as compared to the preceding background rate in the same crustal volume,

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judged significant by some clearly defined standard. The rate decrease takes place within part, or all, of the source volume of the subsequent mainshock, and it extends up to the time of the mainshock, or may be separated from it by a relatively short period of increased seismicity rate. Usually the rate decrease is larger than 40%, and takes place in all magnitude bands.

The pioneering work of Mogi (1969) first brought attention to the possibility that reduced seismicity rates may exist in the source volume before some mainshocks. In this summary we included only one of these examples, however, because they were based on data with M> 6 only, because we do not have estimates of the significance of these rate changes, because the rates are mostly contrasted to neighboring volumes which may pro-duce events at different rates, and because questions about catalog homogeneity exist (Perez and Scholz, 1985). The earliest successful prediction was made by Ohtake et al. (1977) for the 1978 Oaxaca (Mg=7.7) earthquake. Since that publication almost ten years have past without additional successes. One exception may have been the case of the 1980 Mammoth Lake earthquakes (Van Wormer and Ryall, 1980). However, during May 1986 seismic quiescence was brought into focus again: In the Aleutian islands an MC -7,7 earthquake fulfilled parts of the prediction by Kisslinger et al. (1985), and on the San Andreas fault an Mr =4.7 rupture fulfilled a prediction by Wyss and Burford (1985). Because of these advances it is time to review the existing cases of precursory quiescence, although we feel that the data set is too small to arrive at a final under-standing of the phenomenon.

The Method of Detecting Seismic Quiescence

A short discussion of the identification of quiescence is necessary because it influ-enced our selection of high quality data. During the last few years we have become increasingly aware of artificial rate changes in the catalogs (e.g. Figure 6 in Habermann, 1981b; Figure 1 in Habermann, 1982a; Habermann, 1983; 1986a; Figure 1 in Wyss et al., 1984; Figures 6 and 7 in Habermann and Wyss, 1984a). In order to make a convincing case for quiescence one has to pay careful attention to this problem. Another somewhat unexpected fact is that strong seismicity rate variations exist along strike of all seismic zones we have studied (eg. Figure 8 in Wyss et al., 1984; Figures 1 through 6 in Haber-mann et al., 1986). These spatial variations make it difficult to compare rates between neighboring volumes in many areas. Below we give a brief description of what we con-sider to be steps necessary in defining a quiescence anomaly.

First some evaluation of the catalog homogeneity must be made to be certain that changes in the reported seismicity rates are not artificial. A lower magnitude cutoff (Mmin) can be used to eliminate events which are not homogeneously reported. It is desirable to choose Mmin as low as possible, because quiescence can be defined better with more data. An example of the interference of a reporting change with a real ano-maly is provided by the case of the data before the 1978 Oaxaca earthquake. In Figure 3 of Ohtake et al. (1977) one can clearly see that many events with m^ < 4.5 were reported prior to 1968, while none were reported after this date. Habermann (Figure 6 in 198Ib, 1982a) has shown that in the world catalog, and specifically in Mexico, report-ing of small events changed at that time. Luckily, precursory quiescence anomalies are often so strong that interference by artificial changes does not negate them. The Oaxaca 1978 precursor still existed in the data with a lowcut applied (Figure 11 of Habermann, 1981b), but the significance of the anomaly was reduced.

The main danger in predicting earthquakes based on quiescence is that artificial rate changes present in the catalog may be interpreted erroneously as precursors to

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future mainshocks. Examples of strong artificial rate changes which have this potential are discussed by Habermann (1982a, 1982b, 1983, 1986a, 1986b) and Habermann and Wyss (1984, Figures 6 and 7). We found that one does not necessarily obtain a homo-geneous catalog by simply omitting the small earthquakes below Mmin (Habermann and Wyss, 1984) because magnitude shifts also create artificial rate changes (eg. Habermann, 1986a,b). Before searching for quiescence in a catalog we search for rate changes which might have been caused by magnitude shifts and we attempt to correct the catalog mag-nitudes (eg. Habermann, 1986 a,b).

Anomalous seismicity rates cannot be recognized unless the background or normal rate is well defined (Habermann and Wyss, 1984). We have found that remarkably con-stant background rates exist in most of the regions we have examined (eg. Figures 4, 5 in Wyss, 1986; Figures 5, 7 in Wyss and Habermann, 1986), if dependent events are removed from consideration. Identifying these events is usually easy in teleseismic data sets, but can be difficult in local ones. In some tectonic areas large mainshocks occur but the microearthquake activity is very low, even nonexistent. In such areas back-ground rates cannot be established, and hence seismic quiescence cannot be used as an earthquake prediction tool.

The quantitative evaluation of the rate changes is a requirement for taking a reported change seriously. The amount of change is interesting but it is not the only fac-tor which determines the significance of a precursor. The variance of the mean rates also has to be taken into account. If the entire background period contains 10 events and the proposed anomalous period contains 3 events the decrease is 70%, but few people would actually believe that the amount of information was sufficient to define a rate change. Therefore, we have proposed that some statistical means be used to estimate the signifi-cance of the rate change. Based on the significance observed for precursory quiescence and for false alarms we then may be able to empirically derive a significance threshold for declaring alarms without generating false alarms. With this aim in mind we have defined false alarms as rate decreases within the study data set which exceed the target anomaly in significance, but which are not followed by a mainshock in or near the volume in question.

The usefulness of a precursor for actual predictions depends on the number of false alarms generated. To evaluate the false alarm rate we have searched systematically for periods of quiescence with z-values exceeding the target z-value (eg. Habermann, 1981a,b; Wyss et al., 1984; Wyss and Burford, 1985; Wyss and Habermann, 1986). The search was usually conducted in a seismic zone about ten times longer than the target volume (the source volume of the mainshock). The searched zone was divided into abut-ting volumes equal in dimensions to the target volume, and originating at random coor-dinates. In addition a second set of volumes of the same size but overlapping 50% of adjacent volumes of the first set was searched. In all of these volumes the mean rate within a moving time window of the anomaly length was compared to the background rate by the z-test. The resulting z-values can then be compared to the target z-value (eg. Wyss and Burford, 1986; Wyss and Habermann, 1986). If the target z-value is larger than all others then the anomaly can be recognized without generating false alarms. If the z-values in one window or in a group of consecutive windows in one volume exceeds the target z-value then we have found one false alarm.

With time the method to define quiescence and false alarms has improved and is still improving. For this reason some of the older data sets included in Table 1 do not conform to the rigor which we demand from more recent data analyses.

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Characteristics of Seismic Quiescence

The cases of precursory quiescence for which we have detailed information are sum-marized in Table 1. For most of these we know the significance of the rate change and the amount of it, the spatial and temporal extent have been defined quantitatively and the surrounding area has been searched for false alarms. In addition evaluations of the catalog homogeneity was made.
The amount of rate decrease is larger than 40% and can reach 90% (Table 1). Therefore periods of seismic quiescence are often very obvious and resilient to different approaches, or even errors in analysis. In four cases enough events were reported so that it was possible to measure rate changes separately in subregions of the rupture volumes (cases no. 4, 9, 11, 16 in Table l). Strong spatial variations in the amount of rate decrease occurred in all of these cases. The rupture initiation points were located in volumes of little or no change, surrounded by volumes of substantial rate decreases (Wyss et al., 1981a; Wyss, 1986; Kisslinger, 1986; Wyss and Habermann, 1986).We have interpreted this observation as indicating heterogeneous stress (strength) distribution in the source volume, with major asperities in which the mainshocks nucleate showing no quiescence (Wyss et al., 1981b).

The variance of the mean rate during the anomalous time is remarkably low in most cases. This is especially clear in those examples with large numbers of earthquakes (eg. Figures 4 and 8 in Wyss, 1986; Figures 2 and 4 in Kisslinger, 1986). This constant rate of earthquake generation during the precursor time supports our assumption that the anomalies reflect a change of process in the rupture zone. First some process produces events at a constant background rate, then a change takes place and events are pro-duced again at a constant but reduced rate.

The spatial extent of the quiescence anomalies is seldom sharply defined. In telese-ismic data the earthquake density per volume is too low to allow sharp definition of the edges. In some of the local data sets not enough earthquakes occur beyond the end of the main rupture. In most data sets the anomaly extent seems to approximately equal the source volume. In few cases it appears that the quiescence volume may be larger than the source volume (no. 8 in Table 1, Figure 5 in Wyss et al., 1984). Where the source volume was subdivided the volume of quiescence was found to be smaller than the total source volume, but the periphery of both seemed to approximately coincide. Thus, if a quiescence anomaly is defined in real time and an alarm is issued, one has no reason-able alternative but to assume that the future rupture length will equal approximately the length of the quiescence. Wyss and Burford (1986) were successful with this assump-tion, and Kisslinger et al. (1985) would have come closer to estimating the magnitude of the May 7, 1986 Andreanof islands earthquake correctly if they had made the same assumption.

The onset time of quiescence is defined sharply in most cases. The data sets with
the most numerous earthquakes again demonstrate this most clearly (eg. Figure 4, 5 and 8 in Wyss (1986), Figures 2 and 4 in Kisslinger (1986), Figures 6 and 10 in Wyss and Burford (1985)). In some of the cases where subvolumes of the source were analyzed separately it seems that quiescence appears at slightly different times in the different subvolumes (Wyss, 1986; Kisslinger, 1986). However, with so few examples available, one cannot discern a pattern yet for the spatial distribution of onset time.

The duration of quiescence is well defined because the onset time is sharp. In those cases where the quiet period is separated from the mainshock by renewed activity we will use the entire precursor time, from onset of quiescence to mainshock, in the following

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discussion. The durations of quiescence reported (Table 1) range from a little over one to 6.3 years. These limits may be imposed artificially by the data. Because high quality catalogs cover 10 to 20 years only, it may not be possible to define quiescence anomalies of 10 years, even if they existed. The short duration cutoff for anomalies is a function of the reporting rate and the variance from the mean rate. If the reporting rate is very high and has a low variance, then short periods of anomalous rate may be defined. In our experience the shortest periods for which quiescence can be defined by the data we have used without generating false alarms is approximately one year (eg. Wyss and Haber-mann, 1986).

The magnitude band within which the quiescence anomalies reside seems to include
all magnitudes. When only larger background events are used the rate decrease tends to be less significant according to statistical tests. This is so because fewer events are avail-able to define the rate, and hence the variance is large. When all events above Mmin are included we find that the significance of the anomaly is stronger because the variance of the background is smaller.

The false alarm rate is low in data sets with man-made changes accounted for. In spite of fairly extensive searches for false alarms we have not found any, except for those in the Aleutian islands (Habermann, 1981a). The average false alarm rate cannot be accurately estimated as yet, but we expect it to be lower than 50%.

Failures to predict are defined as occurrences of mainshocks without precursory quiescence. A few such cases have been found in the same areas where successes were also reported (Habermann, 1981b; Wyss et al., 1984; Wyss, 1986; Wyss and Burford, 1986). Dividing the number of failures by that of successes in the same areas one arrives at approximately 50% as the average failure rate. Again, this value is likely to change as more information becomes available, and it may also be a function of tectonic setting.

Mechanism of Quiescence

There is not enough evidence available to discriminate between possible mechan-isms which may cause quiescence. As it is most important at this point to learn more about the phenomenon, we have de-emphasized speculations about the mechanism. Below we will briefly outline three obvious possibilities which may cause quiescence in different tectonic environments.

Strain softening may lead to a lowering of the ambient stress in the source volume, and hence every asperity which was near rupture stress and which would have ruptured in the next unit time has a lower probability of rupture after the volume has passed through peak stress. Geodetic observations from the south coast of Hawaii support this model. In the epicentral area of the Kalapana earthquake (M~ =7.2, Nov. 1975, depth = 8 km) the lengths of lines between 3 and 8 km long had been measured since the begin-ning of this century, and it was found that compressive strain of about 4-10 had accu-mulated by 1973. Because of this Swanson et al. (1976) suggested that a mainshock may occur in this area. During one to several years before this mainshock at least one, and probably three, of the geodetic lines lengthened (Figures 16 and 17 in Wyss et al., 198la). The measured strain-drop on the longest of these lines was 3-10~ correspond-ing to a stress-drop of at least 20 bar.

Based on this observation Wyss et al. (1981a,b) proposed that precursory creep on
the near horizontal subsequent mainshock fault plain lowered the ambient stress in the crustal volume above the fault plane. The parts of the source volume in which quiescence did not occur were assumed to be asperities where fault creep did not take place. This

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model assumes that the earthquake population which showed quiescence is distributed in the crustal volume above the fault plane. This is likely to be the case, because most earthquakes in this area have hypocenters between 5 and 8 km depth.

The locked fault model may be more attractive as an explanation for the precursory quiescences observed in the creeping segment of the San Andreas fault, because creep-retardation was observed at the Lewis Ranch creep-meter (Figure 2a in Burford and Schulz, 1985). The onset of the creep-retardation and the quiescence coincided approxi-mately, and the creep-meter was located at the northern edge of the fault-segment along which quiescence was observed before the 31 May 1986 mainshock (Figures 1 and 2 in Wyss and Burford, 1986). In this model we propose that by some unknown means a seg-ment of the creeping fault is locked partially. As a consequence the creep-rate and the seismicity rate are reduced. This model assumes that the counted earthquakes are located on the fault itself, not in the crust surrounding it.

Dilatancy hardening appears to be a less likely mechanism for quiescence because we know of no evidence supporting this explanation. In the contrary, the seismicity rate in the immediate vicinity of the hypocenters, where dilatancy hardening should be most pronounced, does not decrease during the precursor time. In addition the available geo-detic evidence supports the two mechanisms proposed above rather than dilatancy har-dening.

It is not attractive to explain the same phenomenon by different mechanisms (strain softening and locked fault model) in different locations. One would prefer a unify-ing explanation. However, the presently available data do suggest that different mechan-isms may produce quiescence in tectonic settings where the background activity is distri-buted through a crustal volume, and settings where it is located in a narrow fault zone. The ideas proposed here should be refined by quantitative and specific models for case histories. The detailed comparisons of crustal deformation with seismicity data is very important, but difficult to make, because too few geodetic lines and creep meters exist.

Discussion and Conclusions

Progress in learning the facts about precursory quiescence has been rather slow since its discovery almost 20 years ago. However, recent quantitative systematic analysis of the phenomenon (Table 1), and recent successes in predictions based on quiescence strongly suggest that precursory quiescence has occurred and will be useful for predic-tions. The main task now is to establish the characteristics of this precursory parame-ter, and to further refine methods by which correct predictions with a minimal false alarm rate can be made.

The number of high quality quiescence cases is still small (Table l). Too small for instance to answer the important question as to what the precursor time depends on. The idea that precursor times may be a function of the mainshock magnitude is linked to the dilatancy diffusion model (eg. Scholz et al., 1973; Rikitake, 1975). At the moment there is no compelling reason for which one can accept or reject that idea. The quies-cence data in Table 1 come from a variety of tectonic settings, which may influence the precursor time. Plotting these data versus magnitude one finds a weak correlation (Fig-ure 1). The two data points from the creeping segment of the San Andreas fault do not fit the trend of the other data. Their quiescence times are longer than expected for such small mainshocks. If one omits these two points one can find the following least squares fit (minimizing TQ and units in months)

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logTg=0.0 + 0.21M (1)

with a correlation coefficient of 0.7 (dashed line in Figure 1). This relationship is close to the one proposed by Scholz et al. (1973) for M = 6, but it is a factor of 6 below that relation (dotted line in Figure 1) for M=8. The relationship of Rikitake (1975) overesti-mates the quiescence precursor time by an order of magnitude for M = 8 events. In our opinion the existing data are not sufficient to allow a decision whether a quiescence-time versus magnitude relationship as in (l) is valid or not, and whether differences between tectonic settings exist.
Given these uncertainties, the occurrence time of a future mainshock is difficult to estimate based on quiescence alone. We propose that the magnitude should be estimated from the anomaly dimensions, and the precursor duration should be estimated as the average of known quiescence precursor times in the same seismic zone, or in tectonically similar seismic zones.

The false alarm rate and the failure rate are also poorly established at this time. And more fundamentally, questions concerning the homogeneity of the most important earthquake catalogs are not answered, and probably will not be answered for some time. So we find ourselves in a situation where precursory quiescence seems to be the most successful parameter for prediction of mainshocks, but the adequacy of the funda-mental data sets are in question. In addition the most important areas of earthquake hazard, California and Alaska, have not been searched systematically for precursors, false alarms and failures. And the existing data set is so small that we cannot say what the precursor time depends on. Nevertheless, in May of 1986 two earthquakes happened which were predicted based on quiescence. However, the surface has hardly been scratched in the search for precursors to past mainshocks. The data base which we now have (Table 1) could probably be doubled or tripled based on existing seismic catalogs.

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MAGNITUDE

Figure 1: Logarithm of seismic quiescence duration as a function of magnitude (data from Table 1). Circles mark earthquakes from subduction zones, squares indicate intra-plate events, and triangles are data from the San Andreas fault system.
The dashed line is a least squares fit through the data without the values for the
two smallest events. The dotted line is the relationship proposed by Scholz et al. (1973).

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PRECURSORY QUIESCENCE BEFORE THE AUGUST 1982 STONE CANYON, SAN ANDREAS FAULT, EARTHQUAKES

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Abstract

The Stone Canyon earthquake sequence started during August 1982 and lasted for about four months. The sequence contained four mainshocks with M^ 4, each with an aftershock zone about 4 km long. These mainshocks ruptured a segment of the fault approximately 20 km long from the south to the north leaving two gaps, which later produced the M^ = 4.6 mainshocks of January 14, and May 31, 1986.

Precursory seismic quiescence could be identified in: (1) the northernmost 10 km of the aftershock zone which contained three of the mainshocks; and (2) the southern gap in the aftershock zone. The fault segment containing the first mainshock and its aft-ershocks did not show quiescence. This pattern of precursory quiescence is very similar to two cases in Hawaii where the rupture initiation points of the mainshocks (Mg = 7.2 and 6.6 respectively) were located in volumes of constant seismicity rate, surrounded by volumes with pronounced precursory quiescence.

The precursory quiescence before the August 1982 Stone Canyon earthquakes lasted for 76 weeks, amounted to a reduction in rate by about 60%, and could be recog-nized without any false alarms, that is the anomaly was unique within the 60 km study segment of the fault and in the years 1975 through August 1982. Eighteen foreshocks occurred between July 27 and August 07, 1986. We conclude that the August 1982 mainshocks could have been predicted based on seismic quiescence and foreshocks.

Introduction

Recent successful earthquake predictions based on the hypothesis of precursory quiescence added to the earlier success by Ohtake et al. (1977). Kisslinger et al. (1985) observed a decrease of seismicity rate in the area covered by the Adak seismograph net-work. He interpreted this observation as a possible precursor to a large earthquake. Although some of the specific details did not come true, the May 7, 1986 M^ 7.7 Andreanof Islands earthquake essentially fulfilled the prediction (Kisslinger 1986). Along the creeping segment of the San Andreas fault the largest earthquake in four years was predicted correctly one year in advance based on seismic quiescence and fault-creep retardation (Wyss and Burford, 1985). The Stone Canyon shock of May 31. 1986

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ruptured the fault segment specified, measured MT = 4.6 to 4.7 (Mj = 5 was expected), and occurred within the one year time window.

We define precursory quiescence as a highly significant rate decrease (in a restricted segment of a seismogenic zone), which is terminated by a mainshock whose source volume includes all or a major part of the quiescent volume. The amount of seismicity rate decrease is usually about 70%, and the anomaly durations are 1 to several years. The rate decrease should be present in all magnitude bands above the minimum magni-tude useful for analysis, otherwise it is suspected that the anomaly may be artificial.

The expected duration of precursory quiescences is the least well known parameter
in predictions based on quiescence. Thus there is a need to study carefully and in detail as many case histories as possible. The purpose of this paper is to determine whether precursory quiescence existed before the August 1982 earthquake sequence which occurred in the Stone Canyon segment of the San Andreas fault. This sequence was the most significant rupture in the creeping segment of the San Andreas fault since the Bear Valley sequence of 1972/73. The main questions we seek to answer are: Were the August 1982 shocks preceded by quiescence? What relationship did the quiet volume have to the main rupture volume? How long did the precursor last, and what is the false alarm rate associated with this precursor?

Data Selection and Adjustments

The crustal volume for study must be selected larger than the rupture of the mainshock in order to define artificial rate changes if they exist in the catalog. Also, estimates of the anomaly extent depend on comparison of rates within neighboring volumes. For this study we chose a 60 km segment of the San Andreas fault (Figure 1) which contained the mainshocks near its northern end. Farther north along the fault the seismicity rate is too low along most fault segments for rate studies.

Perpendicular to the belt of seismicity along the San Andreas fault we have res-tricted our study to a 5 km wide strip. This width was chosen because it contains the seismicity associated with the main fault strand, but excludes most of the diffuse activity adjacent to the fault, and seismicity located on nearby faults. No earthquakes were excluded on the basis of depth because essentially all hypocenters are located in the top 12 km of the crust.

The detection of seismic quiescence, and its definition in space and time, is best achieved if numerous earthquakes are available for analysis. Therefore we are interested in including the smallest magnitudes reported homogeneously. Although it is desirable to extend the analysis as far back in time as possible, in order to define the background rate, the data before 1969 cannot be used in central California because the reporting rate was much lower. If these data were included the minimum magnitude would have to be raised to an unacceptably high level. Also the Bear Valley earthquake sequence caused a major disturbance of the seismicity budget of this fault segment in 1972/73. Before this time the rates were rather low, after it they were high, and remained high, well into 1974. To avoid this disturbance in our analysis we chose a starting date for the data set in the second half of 1974.

The magnitudes assigned to earthquakes in most catalogs are not stable over long periods of time (Habermann, 1983; Habermann and Wyss, 1984; Perez and Scholz, 1984; Habermann, 1986a,b). This is not surprising because seismograph stations are esta-blished and removed, and algorithms for estimating magnitudes as well as analysts change over the years. For our purpose even small artificial changes in the average mag-nitude assigned to the recorded events can be a problem. For measuring the seismicity

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rate we count the earthquakes reported per unit time with M > Mmin. Earthquakes smaller than Mmin are not counted because they are not reported homogeneously. Sup-pose that at some time, t, new instruments were introduced. This might have the effect of reducing, in the average, the magnitudes by DM = -0.1, in spite of all efforts to com-pensate correctly for the change. (Note that we do not propose that individual magni-tudes are reliable to 0.1 units. Instead we suggest that after the instrumental change a new magnitude scale was established which differs from the old one by DM). Thus, after the change, earthquakes are no longer counted which would previously have fallen in the magnitude band Mmin -f- DM, and an apparent seismicity rate decrease would be detected at time t. Below we explain how we attempt to identify and correct for such artificial magnitude shifts.

The problem is to detect and define the time and amount of magnitude shifts, as well as the time and amount of changes in reporting of small earthquakes. Often the two occur at the same time and are traceable to the same adjustment in the network (e.g., Habermann and Wyss, 1984). To find such changes we use the magnitude signa-ture method (Habermann, 1983; 1986a,b) in the following way. We search the catalog of the entire fault segment for rate changes, assuming that these changes can be classified into three types: (l) Rate changes due to changes in the process producing the earth-quakes, for example a change in the state of stress. Aftershock sequences and precursory quiescence fall in this category. An important assumed characteristic of this type of change is that it occurs in all magnitude bands. (2) Rate changes restricted to small events will be interpreted as due to changes in the reporting capability. (3) Rate changes which are present with opposite polarities in magnitude bands of large and small earthquakes, can often be successfully modeled as magnitude shifts (for details see Habermann, 1986a,b).

The search for times of rate changes is done using the algorithm described by Habermann, (1983) together with visual inspection of cumulative seismicity curves. A magnitude signature is made for each significant rate change detected, and the observed signature is modeled by a synthetic signature containing a magnitude shift, a reporting change for small events, or both. Minimizing the residual between the observed and the computed magnitude signatures we then arrive at Mmin and DM which we propose as the most likely quantities for these parameters.
Some of the rate changes detected and modeled in this fashion may not warrant correction of the catalog, because they may have lasted only a short time, because they may be due to aftershock sequences, or because they may have a small DM. Therefore all potential changes were reviewed after the first examination and the stronger ones were retained. Then new magnitude signatures were computed comparing the rates between times of these changes only. After the new magnitude signatures had been modeled, the resulting DMs were used as preliminary corrections to the catalog. Subse-quent examination of the corrected cumulative seismicity curves for individual 10 km sections of the fault showed that the results were not uniformly satisfactory. The correc-tions were different in their effectiveness in the northern and southern parts of the study area. Hence we divided our study area at latitude 36.63 ° N, and repeated the whole pro-cess of computing and modelling magnitude signatures separately for the north and the south. The final magnitude shifts are discussed below, and corrections for these shifts were applied to the catalog before further analysis.

In selecting magnitude shifts important enough to make a correction in the catalog
we were conservative in the sense that we aimed at a minimum number of corrections. The starting date of the data set was moved up to 22 November 1974 in order to avoid a

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correction at that time. Between this starting date and the time of the mainshock (August 1982) three magnitude corrections were necessary in both regions. Figures 2 and 3 show the magnitude signatures for these changes and the synthetic signatures modeling the changes.

The method of computing synthetic magnitude signatures (Habermann, 1986a,b) is the following: A trial DM is added to each M of earthquakes which make up the back-ground sample period (for example, November 1974 to April 1976 in Figure 2A). Com-paring the rates in all magnitude bands of this shifted data set with the original data set, one then obtains a magnitude signature which results if the magnitude scale has been changed by the amount of the shift. The comparison with the observed magnitude signature is made for a number of different assumed shifts, until a minimum is found in the residuals between observed and synthetic signatures.

The magnitude signature in Figure 2A is typical for a simple magnitude shift. The observed rate changes in the bands of larger magnitudes indicate a reporting rate increase (fairly substantial negative z-values), while the small magnitudes show a very strong decrease in reporting (large positive z-values). This pattern can be modeled closely by assuming that a magnitude shift of +0.12 took place in April 1976 (see * in Figure 2A). The agreement between the synthetic and observed signatures could be made almost perfect in the middle of the plot by assuming that the reporting rate of events smaller than 1.4 was reduced by about 10%. We did not include this feature in our synthetic signature in Figure 2A because we were interested in demonstrating the effect of a magnitude shift alone. Although we do not claim that this interpretation is unique, we feel that the modeling is very convincing and simple. No other effect that we could think of, including a b-value change, would produce this magnitude signature.

In February of 1978 an almost equal magnitude shift with the opposite sign occurred, coupled with a decrease in reporting rate of events smaller than 1.6 by about 45% (Figure 2B). The fit of the synthetic magnitude signature to the observed one is again very good. In June 1980 a simple pattern of positive z versus negative z for larger and smaller events respectively suggests a magnitude shift of-0.1 for the southern region (Figure 2C).

In the northern region (Figure 3) the changes occurred at similar times as in the south, except for the first change. Although there are some similarities between north-ern and southern signatures (eg. June 1980, Figures 2C and 3C), there are also strong differences as in 1976 when the shifts had different signs (Figures 2B and 3B). Also for each northern synthetic a substantial reporting rate change for small events was defin-itely needed to achieve a good match (50% increase in 1975 below M = 1.7, 40% decrease in 1978 below M = 2.0 and another 50% decrease below M = 1.3 in 1980, Fig-ure 3A, B, C respectively). The seismicity catalog was corrected assuming that the mag-nitude shifts as modeled in Figures 2 and 3 had taken place. The current magnitude values were used as the norm.

The minimum magnitude of homogeneous reporting fluctuated frequently. All reporting rate changes affected earthquakes below M^ =1.7 only, except for a rate change in mid 1978 in the northern region which may have affected events up to M^ = 2.0. We have nevertheless chosen Mmin = 1.7 because we felt it was preferable to accept the risk of a false alarm in mid 1978 due to artificial data fluctuations, than to omit the data in the magnitude band between 1.7 and 2.0.

The declustering of the catalog was done using Reasenberg's (1985) algorithm after the magnitude corrections were applied, and after the events smaller than Mmin were

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removed from the catalog. It is essential that the corrections, and culling of the catalog precede the declustering, because otherwise the algorithm will label more earthquakes as clustered during times of increased reporting. This is so because during times of increased reporting the interevent times are smaller due to the presence of more earth-quakes per unit time.

Declustering of the catalog is needed before we begin the analysis of seismicity rate, because we hypothesis that there exists a background rate which is approximately con-stant. Aftershock sequences, swarms and doublets need to be removed from the catalog in order to define the background rate. The catalog correction and declustering was done on a data set from an area three times as large as the study area in order to avoid incorrect declustering along the boundaries of the study area.

Method

Based on experience with a number of searches for quiescence before mainshocks (Habermann, 1981a,b; Wyss et al., 1981a; 1984; Habermann and Wyss, 1984; Wyss, 1986) we have adopted the following procedure to identify possible precursory quiescence and false alarms. The source volume of the mainshock is the target volume. We hypothesis that within all or part of it quiescence may have existed for an extended period before the mainshock. For comparison of volumes containing constant seismicity rate with the potentially anomalous ones we divide the study area into abutting volumes approximately equal to the source dimensions (even numbers of the 600-series in Figure 1). The northernmost of these volumes is positioned at a random location near the northern end of the study area. An additional sequence of abutting volumes of the same size is defined such that they overlap the first set of volumes by 50% (odd numbers in the Appendix and the example 619 in Figure 1). Furthermore two sets (even and odd as above) of smaller volumes are defined in order to detect small scale patterns of seismicity rate changes. These volumes were about 5km long. Volumes smaller than this usually do not contain enough earthquakes to permit analysis of the rate. The particular volumes used in this study were defined by Wyss and Burford (1985). Hence they were in no way tailored to fit any aspects of this study, or to optimize any patterns.

The seismicity rate within each of these volumes (31 in all) was examined by inspection of the cumulative number curves, and by the RAS algorithm (Habermann, 1981a,b; Habermann and Wyss, 1984; Wyss and Burford, 1985). This algorithm relies on the standard deviate z-test to evaluate the significance of the difference between two mean rates in adjacent periods. In doing so we assume that a change of process may have changed the mean rate of production at the time separating the two periods.

By inspection of the seismicity rate within the target volume we define a target window approximately equal in length to the possible anomaly duration. Based on the apparently low rates before the mainshock in volume 610 (Figure 4) we used 76 weeks as the window length. The z-value for the comparison of the mean rate in this target win-dow compared to the background mean in volume 610 was z = 3.99. If we find 76 week windows with larger z-values we will consider these occurrences to be false alarms. In the search for false alarms we move the window through the data set of volume 610, stepping one week at a time, always comparing each window-mean to the mean of the expanding background. In this way the function RAS(t) (Figure 4) is generated. The same process is carried out for all the volumes defined (RAS in Figures 5 through 7), and a histogram is made from all the z-values contained in the RAS functions (Figure 12).

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Alternate statistical tests (e.g., Ohtake et al, 1977) are also used to evaluate the anomaly so that we can be sure that the identification of an anomaly does not depend on the statistical test chosen to evaluate it. Since the quiescence anomaly is obvious to the eye (Figures 4, 8), one might argue that the statistical tests are not needed to make a convincing case. However, we wish to evaluate the anomaly quantitatively to confirm that the rate change is unique, and to arrive at an empirical threshold level for z, above which a present day anomaly should be interpreted as a possible precursor.

Seismicity Rate

Representative data sets for the 600 series of volumes are shown in Figure 5. These data show that in most cases the seismicity rate is fairly constant over many years. This confirms the assumption that a background rate exists against which quiescence may be evaluated.

The data from volume 608 were chosen to demonstrate that background rates can
be constant even in volumes of low seismicity. This was the volume with the fewest events of the 600 series, and it abuts the anomalous volume to the northwest. Volume 611 also shows a constant rate over the entire period. It was chosen as an example because it contained the first of the mainshocks which occurred in August of 1982. The seismicity rate in volume 618 is also constant with minor fluctuations. This data set is typical for the southern fault segments.

Figure 5d shows the data set containing the largest deviations from normal, except
for the target volumes. This is the only volume of the 600 series (except for target volumes 610 and 609) in which the RAS function exceeds the dotted line (Figures 4 and 5). The period centered on 1977 in this volume is the strongest contender for a false alarm. The maximum z-value reached is 3.2 which is less than that of the target volume (z = 3.99) and thus the entire 60km fault segment does not contain any false alarms.

In order to define the spatial extent of the anomaly we evaluated the seismicity rate in the 5km long 700 series of volumes. Figures 4 and 6 show that the volumes 725, 719 and 720 contain quiescence. The latter two of these cover the same fault segment as volume 610, while 725 is located further south (Figure 9). The volumes 721 and 723, located between the quiet ones (610 and 725), show constant rates (Figure 7A, B) although they were part of the August 1982 rupture volume.

The data of the 700 series with the strongest candidate for a false alarm came from volume 728 (Figure 7C). The RAS function shows a peak with z = 3.6, this value is larger than that of the target window in 719 which is z = 3.4 and larger than that in 725 which is only 2.8. Figure 7D shows a typical case for the stable seismicity rates gen-erally present in the south of the study area. We conclude that in the small volumes the strongest anomaly is not a precursor. Therefore, the quiescence in volumes 719, 720, and 725 (the target volumes) are not strong enough to be recognized without a false alarm. If we needed to use these small volumes to define the precursor to the August 1982, we would have to allow one false alarm. However, the quiescence precursor is defined in volumes 610 and 609, in which the anomaly is stronger than in any other volumes. Thus, the precursor is defined without generating a false alarm. We find that the August 1982 earthquakes were preceded by a period of quiescence as proposed by Wyss and Buford (1985) based on a preliminary, less detailed analysis.

Main and Aftershocks of the August 1982 Earthquake Sequence

The August 1982 earthquake sequence consisted of four mainshocks with Mr ^ 4.0. Their epicenters were distributed over a thirteen km segment of the fault (Figure 9), and

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they ruptured sub-segments progressing from South to North during a five week interval (Table 1). The clustering algorithm (Reasenberg, 1985) identified 18 events which occurred during the two weeks prior to the first mainshock as clustered with the rest of the earthquake sequence. They may therefore be considered foreshocks. Spatially the foreshocks were not tightly clustered, but spread out over the main 16 km of the subse-quent aftershock zone (Figure 9A).

The clustered events following each of the mainshocks, but preceding the next one, seem to outline newly ruptured segments, as well as the aftershock zone of the previous main shock (Figure 10). The time interval shown in Figure 10 after the last main shock is longer than the others because it extends to the end of the cluster sequence.

The combined aftershock sequences are not continuous but two noticeable gaps exist between them: a small but distinct one is located between the aftershocks of the first and second main shocks. A second gap exists south of the first sequence, separating the southernmost part from the rest of the cluster sequence (Figures 9, 10).
A simplified interpretation of these 1982 aftershock data is the following. In a five week period most of an approximately 16 km segment of the San Andreas fault broke in four segments each about 4 km long, with an additional minor rupture located 5 km further south. The equivalent magnitude for the entire 1982 cluster sequence as calcu-lated by Reasenberg's (1985) algorithm was 4.9, however if the Berkley local magnitudes are used for the main events then the equivalent cluster magnitude is Mr = 5.0.

Correlation of Quiescence with the Mainshocks

The temporal correlation is simple: the quiescence lasted for 76 weeks before and up
to the mainshock sequence (Figures 4, 6, and 8). But the spatial pattern is more com-plex. The first of the 1982 mainshocks, or the rupture initiation, was located in a fault segment which showed no quiescence (volume 723, 611, for cumulative curves see Figures 7b and 5b), while the next three mainshocks, or the rupture extensions, were located in the quiet volume 610 (Figures 9, 10). This pattern is very similar to that observed for precursory quiescence before two Hawaiian mainshocks.

In November 1975 an Mg = 7.2 mainshock ruptured about 45 km along Hawaii 's south cost. An Mr = 5.9 foreshock preceded it by about an hour. The mainshock itself was clearly a multiple event consisting of 6 sub-ruptures of approximately magnitude 6 each (Harvey and Wyss, 1986). Thus the pattern of rupture was similar to the Stone Canyon case except that the rupture length was three times larger, and that the separa-tion of major events was measured in minutes rather than weeks. In the 1975 Hawaii source volume the fault segment containing the initial rupture also did not show quies-cence, while neighboring parts of the source volume became quiet (Wyss et. al., 1981a). Based on this pattern it was proposed that major asperities may not show quiescence, while other parts of the source volume turn quiet because of slip-softening (Wyss et. al., 198Ib). The same pattern of constant seismicity rate near the focal point, surrounded by pronounced quiescence, was also found in the detailed study of the Mg = 6.6 Hawaiian earthquake of November 1983 (Wyss, 1986), and it was also observed by Kiss-linger (1986) in the Aleutian Islands. It may therefore be that this pattern is to be expected for many earthquakes in different tectonic settings.

We conclude that the correlation of the 1982 quiescence and mainshock sequence in time and space strongly suggest that the two events were related physically. We thus propose that the 76 week long, uniquely significant, quiescence was a precursor to the August 1982 earthquake sequence.

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The quiescence in volume 725 is unexplained by the events of 1982. The lack of aftershocks in this volume suggests that it did not rupture. It turns out that the 1986 Stone Canyon earthquakes of January 14 and May 31 (Mr = 4.6) were located in the two aftershock gaps of the 1982 sequence. Figure 11 shows two weeks of aftershocks for the 1986 earthquakes and their relationship to the volumes 610 and 725. The January 1986 aftershock sequence fills exactly the southern gap left by the 1982 sequence. This mainshock was not predicted because it was not preceded by quiescence that lasted through the time at which Wyss and Burford (1985) made their analysis of the seismi-city rate of this area. Based on the current simple hypothesis of precursory quiescence one would have to reject the idea that the quiescence in volume 725 (Figure 4) was causally connected to the January 1986 mainshock. However, the strong spatial correla-tion of the two phenomena suggests that the two may have been connected, and that for some unknown reason the mainshock which should have happened as part of the August 1982 sequence was delayed until January 1986.

The 31 May 1986 mainshock on the other hand did show precursory quiescence on
the basis of which it was predicted (Wyss and Burford, 1985). This shock was located in the central gap of the 1982 aftershock sequence, however its extent appears to have been larger than the gap. About 60% of the May 1986 aftershock area was located in the fault segment which was not quiet before the 1982 sequence (Figure 11). Most of volume 611 was filled by the May 1986 rupture. It appears that this volume was still capable of precursory quiescence in 1984/85 because it did not turn quiet in 1981/82.
Although these patterns are interesting we do not claim to understand them at the present, nor are we certain that they are generally valid. If more such cases can be defined by high quality data sets like the present, the Hawaiian and the Aleutian cases mentioned above, then one may be able to propose a general phenomenological and phy-sical model for these patterns.

Interevent Times

An alternative measure of rate changes can be obtained by comparing interevent times AT (Udias and Rice, 1975). The advantage of this parameter is that one does not have to introduce an arbitrary sample interval length. The disadvantage is that during quiescence there are few interevent times (zero in the most quiet case) for comparison with the background interevent time distribution.
In the present case in volume 610, we found that for the background period the mean was AT=236± 247 hours, while the anomalous period had a mean of AT=439± 390 hours. The distribution of AT during background time showed a clear peak at 4 hours with a slow decay towards larger values, while the distribution during the anomalous period did not show a peak and short interevent times were rare. Thus, the interevent time analysis confirms our finding that a significantly reduced seismicity rate preceded the August 1982 mainshocks.

Discussion

The background seismicity rate along the study segment on the San Andreas fault
is fairly constant, allowing quantitative evaluation of rate changes. The smallest fault segments which contain enough earthquakes per unit time for such studies are 5 km if Mmin = 1.7 is used. In future studies it may be possible to resolve more detail of seismi-city patterns if post-1980 data with a lower Mmin can be used.

The August 1982 mainshock sequence was the largest rupture of the creeping seg-ment of the San Andreas fault in the last 14 years. It ruptured a segment 16 to 20 km

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long (Figure 9 and 10). However, the rupture was not continuous. Gaps remained between fault segments of dense aftershock activity. The northern two thirds of the August earthquake sequence filled a volume (610) in which seismic quiescence existed during 76 weeks before the first mainshock. This quiescence was the most significant reduction in rate in any of the volumes studied in a 76 week period (Figure 12). Thus, if the alarm threshold is set at z = 3.2 then the anomaly can be identified without any false alarms.

The spatial and temporal correlation of the quiescence anomaly in volume 610 with
the August 1982 earthquakes is very close indeed. Therefore there is no reasonable alter-native to the interpretation that this quiescence was a precursor to the August 1982 earthquake sequence. The quiescence in volume 725, however, occurred along a fault seg-ment where only few 1982 aftershocks were located (Figures 9 and 10), but were the fault ruptured in a mainshock in January 1986 and its statistical significance was not very high. The fact that the two 1986 mainshocks were located in the aftershock gaps of 1982 suggests that the 1982 ruptures set up the 1986 ones. Further we note the facts that the volume which did not show quiescence before 1982, did show precursory quies-cence before the May 1986 rupture (Wyss and Burford, 1986), while volume 725 showed quiescence in 1982, but not immediately before the January 1986 shock. A model to explain the cause of quiescence along this fault segment will have to consider these facts. In this paper however, we restrict ourselves to define quantitatively the details of precur-sory quiescence before the August 1982 mainshocks.

There exist four cases of precursory quiescence in which the seismicity rate per unit volume was high enough to allow separate study of subsegments of the source. These are the Kalapana (Hawaii, November 1975, Mg = 7.2; Wyss et al. 1981a), the Kaoiki (Hawaii, 16 November 1983, Mg = 6.6; Wyss, 1986), the Andoeanoff Islands (Aleutians, 7 May 1986, M = 7.7; Kisslinger, 1986) and the present case. In all of these the initial rupture occurred in a volume of constant seismicity rate, surrounded by volumes con-taining quiescence. This suggests that this may be a general pattern in which major asperities do not turn quiet while the rest of the source does.

The duration of the precursory quiescence before the August 1982 rupture (76 weeks) was short compared to the 154 week quiescence before the 31 May 1986 Stone Canyon earthquake (Wyss and Buford, 1986). Especially when one considers that the August 1982 and May 1986 ruptures extended over about 16 and 7 km respectively. This contradicts the expectation that larger earthquakes might have larger precursor times (eg. Scholz et al. 1973). Further evidence against the latter expectation comes from the comparison with quiescence times before Mg > 7 events which range from 2 to 4 years. Based on the limited data available it thus appears that the quiescence precur-sor time is not a strong function of mainshock size (Wyss and Habermann, 1986). Whether it is a function of tectonic environment cannot be established with the few reli-able data points existing now. This is one of the reasons why it is important to study quantitatively, and in detail, changes of seismicity rate before as many mainshocks as possible.

The example of the search for the August 1982 precursory quiescence confirmed again that the statistical tools we have developed and the method to map volumes of quiescence works well. The large z-value for the 76 week anomaly in volume 610, z = 3.99 implies that the two means are different at a confidence level of 99.6%. If we assume instead that the earthquake process is Poissonian as Ohtake et al. (1977) did, then we estimate the probability to be 99.9%, and with the assumption of a binomial distribution

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(Habermann, 1986a) we estimate p = 99.9%. Clearly all these models lead to the same answer, namely that it is most likely that a reduced rate existed during 76 weeks before the 1982 mainshocks in volume 610.

In the present analysis we have chosen not to optimize the anomaly volume by adjusting the boundaries of the randomly defined volumes. This was done to avoid the criticism that we bias the estimate of the anomaly significance by optimizing the volume. However, we are not convinced that optimization would have introduced a significant bias. We believe that this is a minor point and that it does not really matter whether we rely on analysis in overlapping 5 km fault segments, or whether we design additional spe-cial volumes. We estimate that the boundaries of quiet volumes cannot be defined to within better than 2 to 5 km, because epicenter errors and gradients in rate along the fault trace would diffuse the sharpness of the boundary even if a sharp boundary of quiescence existed.

In this part of the San Andreas fault quiescences within 10 km segments can be recognized as highly significant after their duration exceeds about one year. Thus it is possible to issue a prediction before mainshocks in the Mr = 5 ± 0.5 range as it was successfully done for the 31 May 1986 earthquake (Wyss and Burford, 1986). Apparently not all mainshocks in this range are predictable here, unless the quiescence hypothesis can be refined for this area. The January 1986 Bear Valley quake was not predicted by Wyss and Burford (1985) because it was not preceded by quiescence during 1985. Instead a period of quiescence had already existed prior to August 1982.

We conclude that the August 1982 earthquake sequence in Stone Canyon was pre-ceded by precursory quiescence of 76 week duration. The rate decrease was about 60% and occurred in parts of the source volume. Complex patterns of quiescence may furnish clues concerning the local fault properties, but they will also make it more difficult to correctly interpret quiescence for earthquake prediction purposes.

Acknowledgments

We thank R. Lester for supplying the seismicity catalog data, and D. Gillard for the code computing interevent times. This work was supported by U.S. Geological Sur-vey grant no. 14-08-0001-G1089.

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Table 1. Mainshocks of the August 1982 Stone Canyon Sequence

Date

10 Aug 82
11 Aug 82
31 Aug 82
24 Sep 82

Hour

02
07
03
08

ML (BRK)

4.5
4.6
4.0
4.0

Latitude,N

36.58
36.63
36.64
36.66

Longitude, W

121.23
121.30
121.31
121.33

Table 2. Characteristics of the Precursory Quiescence to the August 1982 Mainshocks

Quiescence

Start

Jan 1981

Duration

76 weeks

Seismicity Rate

Background

0.29± .53

Anomaly

O.lli .31

Rate

Decrease

62%

ML

Equivalent

5.0

Length

Anomaly

10km

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SAN ANDREAS FAULT SEGMENT

36.7

36.3

VOLUME 608

VOLUME 610

VOLUME 619

121.4 LONGITUDE 121.0

Figure 1. Map of the study area with epicenters along the fault zone for earthquakes 1.7 for the period January 1981 to July 1982. The polygons outlined are the even numbered ones of the 600 series. Polygon 619 is shown by dotted lines indicating the position of the odd members of the 600 series.

550

100

CD

2

z> T

20

I I I I I

- SAN ANDREAS FAULT

STONE CANYON

VOLUME 610

130

110

20

- 2.6
I I I

SAN ANDREAS FAULT

STONE CANYON

VOLUME 725

M > 1.7

2.6

- 2.6

75 76 TIME (YEARS) 80 81 75 76 TIME (YEARS) 80 81

Figure 4. Cumulative number of earthquakes as a function of time up to the August 1982
mainshock sequence at Stone Canyon. The volumes are defined in Figure 9 and in
the Appendix. The RAS function shows the z-values resulting from rate comparis-
ons of 76 week windows compared to an expanding background moved through
the data at one week steps.

553

50

40

10

1 I I I

SAN ANDREAS FAULT

VOLUME 608

M > 1.8

2.6

- 2.6

120

100 -

75 76 TIME (YEARS) 80 81

SAN ANDREAS FAULT

STONE CANYON

VOLUME 611

M > 1.8

- 2.6

81

- SAN ANDREAS FAULT

BITTERWATER VALLEY

VOLUME 618

M > 1.8

- 2.6

180

160

75 76 TIME (YEARS) 80 81

- 2.6

75 76 TIME (YEARS) 80 81

Figure 5. Cumulative number of earthquakes as a function of time arid RAS functions for 76 weeks moving windows as in Figure 4 for four selected volumes of the 600 series.

554

SAN ANDREAS FAULT

STONE CANYON

VOLUME 719

M > 1.7

90

- 2.6

70

20

I I I I

SAN ANDREAS FAULT

STONE CANYON

VOLUME 720

2.6

- 2.6

75 76 TIME (YEARS) 80 81 75 76 TIME (YEARS) 80 81

Figure 6. Cumulative number of earthquakes as a function of time and RAS functions for two volumes of the 700 series with precursory quiescence before the August 1982 mainshock sequence which occurred at the right edge of the graph. Same as Fig-
ure 4.

555

100

80

20

2.6

- 2.6

SAN ANDREAS FAULT
STONE CANYON
VOLUME 723
M > 1.7

- 2.6

75 76 TIME (YEARS) 80 81 75 76 TIME (YEARS) 80 81

90

70

20

SAN ANDREAS FAULT
VOLUME 728
M > 1.7

2.6

- 2.6

SAN ANDREAS FAULT
VOLUME 733
M > 1.7

- 2.6

75 76 TIME (YEARS) 80 81 75 76 TIME (YEARS) 80 81

Figure 7. Same as Figure 6 for selected volumes of the 700 series.
556

500

400

100

r- SAN ANDREAS FAULT
M > 1.7

- CONSTANT VOLUME

STONE CANYON
QUIET VOLUME

I I I I I I

2.6

- 2.6

75 77 TIME (YEARS) 83 85

Figure 8. Same as Figure 4 but extended in time past 1982 for the quiet volume (610 plus
725) and the rest of the study area (606 through 624 minus 610).

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36.7

36.5 -

STONE CANYON
CALIFORNIA

610

725

270782 TO 100882

121.4 LONGITUDE 121.2

36.7

36.5

STONE CANYON
CALIFORNIA

M = 4.3

100882 TO 211182

I___________________I
121.4 LONGITUDE 121.2

Figure 9. Maps of the main and fore-shock epicenters (A), and of the entire cluster sequence
of August through November 1982 (B). Separate symbols mark the events in the
four time periods of Figure 10. Volumes 610 and 725 contained precursory quies-
cence. The temporal sequence of mainshocks is labeled by numbers 1 to 4.

36.7

36.5

36.7

36.5

610

STONE CANYON
CALIFORNIA

725

M = 4.3

02H 100882 TO 06H 110882

J_______________I

36.7 -

36.5

121.4 LONGITUDE 121.2

STONE CANYON
CALIFORNIA

M = 4.5

07H 110882 TO 02H 310882

121.4 LONGITUDE 121.2

I
STONE CANYON
CALIFORNIA

M = 4.1

03H 310882 TO 07H 240982

36.7 -

36.5

121.4 LONGITUDE 121.2

STONE CANYON
CALIFORNIA

M = 4.3

08H 240982 TO 211182

121.4 LONGITUDE 121.2

Figure 10. Aftershock epicenter maps for the periods following the four mainshocks of the August 1982 sequence. Volumes 610 and 725 mark the areas of quiescence. ,.,-g

STONE CANYON CALIFORNIA

36.7

31 MAY 1986
M = 4.6

610

14 JAN 1986
M = 4.6

725

121.4' LONGITUDE 121.2

Figure 11. Epicenter map for the two week aftershocks of the 1986 Stone Canyon mainshocks. Volumes 610 and 725 were the areas of quiescence before the 1982 earthquake sequence. The volume of precursory quiescence before the May 1986 mainshock coincided exactly with its aftershock volume (Wyss and Burford, 1986).

160 I I j I I I M I I M M I I M I III A I I I I I I I I I M I I I I I I

_ STONE CANYON , II I I WINDOW 76 _

Figure 12. Histogram of z-values resulting from all rate comparisons of 76 week windows with the background rate. Solid line for M> 1.7, dashed line for M^ 2.0. The values
for the target windows for the two magnitude bands are z(1.8) = 3.99, z(2.0) =
3.22.

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Two Categories of Earthquake Precursors, Physical and Tectonic, and

Their Roles in Intermediate-Term Earthquake Prediction

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Abstract. I suggest that earthquake precursors can be divided into two categories, physical
and tectonic. Physical precursors are direct or indirect indications of initiation or progression of (irreversible) rupture-generating process within the preparation zone of a forthcoming earthquake. Tectonic precursors are manifestations of tectonic movements outside the preparation zone of the impending earthquake as links in a chain of particular local tectonism in each individual area preceding the earthquake.
Almost all intermediate-term, short-term and immediate precursors of various disciplines within the source regions of main shocks are considered physical ones. Some precursory crustal deformations around the source regions are, however, possibly tectonic precursors because they may be caused by episodic plate motions in the neighboring regions of the fault segments to break. A possible example of this kind of phenomena is the anomalous crustal uplift in the Izu Peninsula, Japan, before the 1978 Izu-Oshima earthquake of Ms 6.8. Some of changes in seismicity patterns in wide areas surrounding source regions also seem to be tectonic precursors because they were probably brought about by particular tectonic setting of each region. A typical example is a so-called doughnut pattern before the 1923 Kanto, Japan, earthquake of Ms 8.2. Although most studies on earthquake precursors so far seem to have been regarding implicitly all precursory phenomena observed as physical ones, the two categories should be discriminated carefully when statistical analyses or physical modelings are carried out based on
reported precursory phenomena. In active plate boundary zones, where practical strategy for earthquake prediction may well be different from that in intraplate regions, tectonic precursors can
be powerful additional tools for intermediate-term earthquake prediction.

1. Introduction

In short-temi (weeks to days) and intermediate-term (years to months) earthquake pre-diction precursory phenomena are the most important matter. Therefore, almost all efforts in earthquake prediction research concerning these time ranges have been concentrated on observing earthquake precursors, obtaining general laws on the way of their appearance, constructing physical models to explain such general laws, and, finally, elucidating the mechanism of emergence of precursory phenomena.
It seems that so far almost any precursor observed has been used in such investigations. However, since large-scale seismic faultings which can be objects of earthquake prediction efforts are not only pure physical phenomena of rock failure but also radical manifestations of tectonic movements in the regions concerned, so-called earthquake precursors probably contain various types of phenomena some of which may be inadequate to be referred to in statistical analyses and physical modelings of generalizing purpose.
In this paper I suggest that earthquake precursors are divided into two categories, which
I call "physical" and "tectonic", and emphasize the importance of discriminating these two categories from each other in earthquake prediction research, by referring to precursory phenomena to Japanese earthquakes. I also discuss their roles in intermediate-term earthquake prediction.

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2. Physical Precursor and Tectonic Precursor

In the study of earthquake precursors we have been usually imagining, I think, a crustal domain associated with a large-scale earthquake rupture (preparation zone of an earthquake), which is implicitly a huge analog of a rock specimen in a laboratory fracture experiment. Some investigators may imagine as such a crustal domain a so-called earthquake source region which consists of a large fault segment with 2-dimensional heterogeneity (asperities or barriers) and surrounding rock volume where the strain energy to drive the final rupture is stored, while some others may include broader environs of the source region that have 3- dimensional structural heterogeneity as illustrated in Fig. l(a). Such a crustal domain is considered to be subjected to a uniform tectonic stress loading just as in a laboratory experiment.
Then, we can expect, at a certain stress level, some kind of rupture-generating process such as changes in rock property or structural heterogeneity (dilatancy, closure of cracks, fluid flow, etc.), rupture of weaker asperities, instability of the fault (preslip), and so on to start within this domain. They will become irreversible from a certain stage, and finally the main rupture along the fault plane takes place.
What I call physical precursor is a direct or indirect indication of initiation or progression
of such a rupture-generating process within the crustal domain of a forthcoming earthquake rupture.
Meanwhile, in actual seismic belts on the earth such a crustal domain associated with an earthquake rupture is placed within a local tectonic regime of each individual area in which a few or many major tectonic elements are mechanically interconnected as Fig. l(b) shows. In this tectonic regime, microscopically, a large earthquake is, although drastic, just one step in a chain of local tectonic movement that progresses step by step responding to the local tectonic stress or the motion of lithospheric plates.
What I call tectonic precursor is manifestation of such tectonic movement outside the crustal domain of a forthcoming earthquake as a link in a chain of particular local tectonism preceding the earthquake. So, the physical meaning of this category of precursors is essentially different from that of the previous category.
Almost all intermediate-term, short-term and immediate precursors of various disciplines within the source regions of main shocks are considered physical ones. However, some precursory crustal deformations around the source regions are possibly tectonic precursors because they may be caused by episodic plate motions in the regions surrounding the fault segments to break, which are triggers or fore-steps for the coming earthquakes to occur. Some of changes in seismicity patterns in wide areas surrounding the source regions also seem to be tectonic precursors because they were probably brought about by particular tectonic setting of each region. Although most tectonic precursors are considered to act as triggering factors for main earthquakes by accelerating stress loading, some may be just parallel responses to the same tectonic stress as responsible for the main shocks.
Since most studies on earthquake precursors so far seem to have regarded implicitly all precursory phenomena observed as physical ones, the concept of physical precursor is expected to be easy to understand. So, in the next chapter I will concentrate on showing examples of tectonic precursors. For demonstration of tectonic precursors Japan is one of the best places.

3. Examples of Tectonic Precursors in Japan

3.1. The 1978 Izu-Oshima earthquake
Figure 2 shows some typical precursors to the 1978 Jan. 14 Izu-Oshima earthquake of
Ms 6.8 which took place on the Pacific coast of central Japan. The earthquake was a vertical right-lateral strike-slip faulting [e.g., SHIMAZAKI and SOMERVILLE, 1979] as shown in the figure. At Funabara, Fn, the water level of a deep well began to fall down remarkably about

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one month before the earthquake [YAMAGUCHI and ODAKA, 1978]. At Nakaizu, Ni, both intermediate-term and short-term precursory changes in radon concentration of groundwater were observed [WAKITA, 1981]. And at Irozaki, Ir, a precursory change in crustal volumetric strain of about one month leading time was observed [JAPAN METEOROLOGICAL AGENCY, 1978]. All these stations are located within 30 km from the main fault.
The Izu Peninsula is situated at the northern tip of the Philippine Sea plate and is colliding with the Japanese main island to its north (Fig. 4). The Philippine Sea plate is subducting beneath northeast Japan at the Sagami trough and beneath southwest Japan at the Suruga trough with bringing about repeated large interplate earthquakes along these troughs. The latest events were in 1854 and in 1923 as shown in the figure. ISHIBASHI [1977, 1978] suggested that inside the Philippine Sea plate (R in Fig. 4) a kind of nascent subduction boundary (West-Sagami-Bay fault) has been developing as a result of the collision to the north. West-Sagami-Bay fault ruptured in the latest during the 1923 Kanto earthquake of Ms 8.2 as a westerly-dipping left-lateral reverse faulting [ISHIBASHI, 1985a]. ISHIBASHI [1978] inferred that the current convergence between the Philippine Sea plate and southwest Japan in this region is accommodated by a dual subduction system consisting of the Suruga trough megathrust and the West-Sagami-Bay fault. The southern half of the Izu Peninsula between them is considered a shear zone composed of northwest-southeast striking right-lateral transform faults (Izu Transform Belt). Thus, the Izu Peninsula is now the plate boundary itself.
Very important phenomena in the recent crustal activity in the Izu Peninsula is that a remarkable crustal uplift has been taking place since 1975 (Figs. 3, 4). S ATO [1981] revealed that four major seismic activities in the Izu region during the period 1975-1980 including the 1978 earthquake were preceded by notable progressions of the uplift with several month durations (Fig. 3). ISHIBASHI [1977] proposed an idea that a creep dislocation in some depths along the southern extension of the West-Sagami-Bay fault caused the uplift, and ISHIBASHI and MATSU'URA [1977], following this idea, obtained a set of optimal fault models propagat-ing deeper for the earlier four stages of the uplift by inversion analysis of geodetic data. SATO's [1981] result stropgly suggests that the aseismic thrusting inferred by ISHIBASHI [1977] continued to occur intermittently causing episodic acceleration of the advance of the Philippine Sea plate which resulted in excess stress loading to some other parts of the Izu region. Therefore, we may well interpret that the precursory phenomena to the 1978 Izu-Oshima earthquake shown in Fig. 2 as well as the precursory crustal uplift shown in Fig. 3 are not indications of the rupture-generating process restricted within the source region of the earthquake, such as dilatancy or preslip, but reflections of crustal movements related to such plate motions, which were triggers or fore-steps of the Izu-Oshima earthquake. Thus, they should be placed in the category of tectonic precursors.
There is an alternative interpretation of the crustal activity in the Izu region that is more prevailing than the above-mentioned one. MOGI [1982], for example, attached importance to the volcanic feature of the Izu Peninsula and attributed the crustal uplift to movement of magma (Fig, 5). But, in this case also, the crustal movement and seismic activity are considered complementary each other, and the precursory phenomena to the 1978 earthquake shown in Figs. 2 and 3 can be regarded as tectonic precursors.

3.2, The 1923 Kanto earthquake
Concerning the seismic activity before the 1923 Kanto earthquake of Ms 8.2, MOGI [1982, 1984] showed that a so-called doughnut pattern can be recognized for the period 1885- 1921 (Fig. 6). SEKIYA [1977], on the other hand, reported an anomalous seismicity of 82 year precursor time. RIKITAKE [1982] pointed out that epicenters of moderate earthquakes tended to move toward the epicenter of the Kanto earthquake during several year period before it. Although these are interesting phenomena, and although OHNAKA [1985] also recognized a seismic sequence of precursory activity, quiescence, and foreshocks before the main shock and explained it by a simple physical model, it should be noted that every moderate or large earthquake before the 1923 Kanto earthquake had its own tectonic meaning.

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As shown in Fig. 7, the Philippine Sea plate is subducting beneath northeast Japan at the Sagami trough, and the Pacific plate is subducting beneath both the Philippine Sea plate and northeast Japan at the Izu-Bonin-Japan trench. Note that the subducted Pacific and Philippine Sea plates are contacting each other under northeast Japan as suggested by the depth contours of the upper surfaces of these two slabs.
Although the main rupture of the 1923 Kanto earthquake took place on the interface between the Philippine Sea plate and northeast Japan along the Sagami trough, the largest aftershock (Ms 7.7) of one day after the main shock occurred at the mechanical plate boundary between the Pacific plate and northeast Japan (region K in Fig. 7) [ISHIBASHI, 1985c]. And a remarkable aftershock activity also took place in the collision zone of the Philippine Sea plate to the north of the Izu Peninsula (Eastern Yamanashi area, region Y in Fig. 7). Thus, the 1923 Kanto earthquake sequence, including the main shock and major aftershocks, was, as a whole, a large-scale elastic rebound of the southernmost part of northeast Japan against both the Philippine Sea and Pacific plates.
As for the earthquakes before the Kanto earthquake (see Fig. 7), the 1909 event (Ms about 7.7), for example, was probably a tear faulting within the subducted Pacific slab [ISHIBASHI, 1985c]. The 1895 (M 7.2) and the 1921 (M 7.0) events are both inferred to have been faultings between the subducted Pacific and Philippine Sea plates based on their estimated focal depths. Shocks in the Eastern Yamanashi area are considered to have been direct indications of plate collision. ISHIBASHI [1976] explained the anomalously high seismic activity in this area prior to the Kanto earthquake by a simple mechanical model. Thus, every moderate or large earthquake in this region before 1923 occurred by its own tectonic meaning in a series of tectonic movement converging to the occurrence of the 1923 Kanto earthquake. Therefore, various precursory phenomena related to the seismic activity before the 1923 Kanto earthquake, as integrations of these individual events, are considered to have been mostly manifestations of movements of the tectonic system surrounding the 1923 rupture domain, rather than indications of physical process toward the final rupture within the preparation zone of the 1923 earthquake. So, they should be placed in the category of tectonic precursors.

4. Discussion and Conclusions

In practical earthquake prediction, if an intermediate-term prediction is meaningfully intermediate at all, it should guarantee that the target earthquake will not occur for a certain months or years with considerably high confidence level. Otherwise, there is no practical difference between intermediate-term and short-term predictions. The best way to satisfy this requirement is to find out some causality laws of rupture-generating process within the preparation zone of earthquakes which assure the irreversibility of the process over the intermediate time range. The proposal of the dilatancy-diffusion model, for example, was one of such trials.
Meanwhile, tectonic precursors to Japanese earthquakes seem to be mostly intermediate-term. This is probably general feature of this category of precursors because, if they act as triggering factors of main earthquakes, it will take some time for tectonic movements around the source region of an impending earthquake to accelerate stress loading to the main fault. Probably, tectonic precursors are predominant to interplate earthquakes because in most plate boundary zones many local tectonic elements are supposed to be mechanically inter-connected to accommodate relative plate motions. Intraplate earthquakes may also be preceded by tectonic precursors, but it seems considerably difficult to recognize some anomalous phenomena as tectonic precursors to a specific forthcoming earthquake. In the first place, seismotectonic studies and specification of expected seismic faultings based on them, themselves, are very difficult in intraplate regions at present.
Therefore, practical strategies for predictions of interplate and intraplate earthquakes may well be different from each other. For the prediction of intraplate earthquakes investigation of 3-dimensional heterogeneous structures through seismological and geophysical methods and

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continuous monitoring of stress or strain state and medium property seem to be essential. On the other hand, for the practical prediction of interplate earthquakes, although the same works as above are indispensable, investigations on seismogenic tectonism including recent evolution of plate accommodation can be effectively utilized as discussed by ISHIBASHI [1982]. In this sense, tectonic precursors can be incorporated in the intermediate-term earthquake prediction in plate boundary zones. Since in a particular local tectonic regime almost the same tectonic precursors are expected to reappear for repeated earthquakes on a given fault segment, tectonic precursors can be still more powerful tools if the local tectonism is minutely investigated and past cases in the recurrence series of the same seismic faulting is well documented and well interpreted. Furthermore, when the space technique comes to be used routinely for contin-uous monitoring of plate motions including small-scale block movements in future, the applicability of tectonic precursors to earthquake prediction will become higher.
ISHIBASHI [1985a] pointed out that each of two antecedents (in 1707 and 1854) of the Tokai earthquake, one of the most important targets of earthquake prediction in Japan, was preceded by a destructive earthquake near Odawara, about 50 km away from the nearest part of the hypothetical Tokai source, with a few year leading time. He interpreted that the Oda-wara earthquake accelerates the plate motion to cause excess stress loading to the nucleation area of the Tokai earthquake in a critical condition. If this is true, the Odawara earthquake is regarded as a tectonic precursor to the Tokai earthquake and expected to play an important role in its intermediate-term prediction, although the Odawara earthquake itself is a big problem of earthquake prediction both scientifically and socially as 86 % of its mean recurrence time has already elapsed [ISHIBASHI, 1985b]. At least its occurrence would be able to make the decision makers on earthquake preparedness activities realize that the probability of the occurrence of the Tokai earthquake within a few years has become very high.
However, it is very difficult to guarantee that the Tokai earthquake does not occur before
the occurrence of the Odawara earthquake with any significant probabilistic evaluation. Thus, tectonic precursors may not be an ideal tool of intermediate-term prediction yet. In the case of interplate earthquakes as well, for strict intermediate-term prediction monitoring of stress state or physical condition in the source region is, of course, essential.
It should be noted that there may be another kind of precursors of intermediate feature between physical and tectonic ones, that is, manifestations of physical process in the prepara-tion zone of a forthcoming earthquake at remote places through regional tectonic movement. KANAMORI [1972] presented an example of such precursors by suggesting that the rise of the earthquake swarm activity in the Wakayama area, about 400 km away from the rupture zone of the 1923 Kanto earthquake west-south westward, about 4 years before this great earthquake had been caused by the excess stress loading of the Philippine Sea plate onto the swarm area resultant from removal of the force opposing the motion of the Philippine Sea plate by a preslip of the Kanto earthquake. For this kind of precursors some physical precursors are expected to appear in the source region correspondingly. In fact, in the case of the 1923 Kanto earthquake, precursory vertical crustal movement suggesting a preslip had been observed for about 10 years before the main shock at Aburatsubo, right above the main fault (see Fig. 7), by tide-gauge observation [e.g., OHNAKA, 1985]. Further study on such inbetween precursors would be helpful to elucidate the nature of so-called sensitive points or
zones.

In conclusion, we should discriminate carefully two main categories of earthquake precursors, physical and tectonic, from each other in earthquake prediction research. Spatial distribution and precursor time of tectonic precursors are suspected to have no general corre-lation with source parameters of following earthquakes because tectonic precursors, being controlled mainly by local tectonism in each region, have essentially no physical implication in the rupture-generating process taking place within preparation zones of earthquakes. There-fore, when statistical analyses or physical modelings based on reported precursory phenomena are carried out aiming at general explanation of earthquake precursors, tectonic precursors must be eliminated carefully. Although the strict intermediate-term earthquake prediction should be based on physical precursors which are indications of irreversible rupture-

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generating process within preparation zones, tectonic precursors can be additional powerful tools especially in active plate boundary zones.

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(a)

(b)

WWW

Fig. 1. Schematic illustrations of (a) a cruslal domain associated with a large-scale earthquake rupture consisting of the source region and its environs and being subjected to a uniform tectonic stress loading (preparation zone of an earthquake), and (b) a local tectonic system composed of several major tectonic elements including the cruslal domain of the earthquake concerned.

Funabara Nakaizu Rn

Water level

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- ID- Yamaguchi & Odaka 1978

ix x xi n i ii rv v vi vii v» ix x M xii | i

1977 1978

Wakita 1981

SEP. OCT. NOV. DEC. | JAM. FEB.
1877 1878

Irozaki V. strain

Ext
t

Nov.

JMA ±y /B

"^^__

Dec. 1977

I

Jan. 1978

Fig. 2. Changes in the water level at Funabara (Fn) [YAMAGUCHI and ODAKA, 1978], in the radon concentration of groundwater at Nakaizu (Ni) [WAKITA, 1981], and in the crustal volumetric strain at Irozaki (Ir) [JMA, 1978] precursory to the 1978 Jan. 14 Izu-Oshima, Japan, earthquake of Ms 6.8. Vertical arrows indicate the occurrence time of the earthquake. The lower left shows the earthquake faulting [e.g. SHIMAZAKI and SOMERVILLE, 1979] and the station localities.

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.IliiiL

20--
cm

10--

Ito

up --Tidal Obs. (Ito)

Leveling (BM. 9340)

BM.9340

o----"

1975 76 77 78 79 80

Fig. 3. Temporal change in the crustal uplift at Ito on the east coast of the Izu Peninsula as revealed by tidal observation and leveling (lower), and the four major seismic activities in and around the peninsula during the period 1975-1980 (upper) [SATO, 1981]. The arrow at the stage 2 indicate the Izu-Oshima earthquake. The inset shows rough locations of the seismic activities with the positions of Ito tidal station and BM 9340.

sw

Japan ^ *

Philippine Sea Plate

Fig. 4. Schematic illustration of the present-day plate motion in the region around the Izu Peninsula. Solid lines outside the Izu Peninsula represent major plate boundary faults. The areas with lateral stripes are rupture zones during the 1854 Ansei-Tokai and the 1923 Kanto earthquakes. R shows rough area of the West-Sagami-Bay fault [ISHIBASHI, 1985a]. The suppled area to the north of the Izu Peninsula is the collision boundary zone. The white arrow indicates the moving direction of the Philippine Sea plate relative to northeast Japan, and solid arrows, directions of relative plate motion at each boundary. In the Izu Peninsula recent earthquake faults including the 1978 Izu-Oshima event are shown by solid lines. The hatched area is a shear zone which ISHIBASHI [1978] called "Izu Transform Belt". Rough extent of the anomalous crustal uplift area as of Feb. 1978 [GSI, 1978] are stippled. Locations of Fn, Ni, Ir, and Ito are also shown (see Figs. 2 and 3).

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Fig. 5. Schematic illustration of the interpretation that movement of magma has caused the crustal uplift and triggered swarm-type and large earthquakes in the Izu region [MOGI, 1982].

1922-'23-4
o'23-9-'24-4

140' 142* E 138' 140* 142* E

Fig. 6. Epicentral distributions of earthquakes of magnitude 6.0 and larger before and after the Kanto, Japan, earthquake of Sept. 1, 1923 (Ms 8.2) [MOGI, 1982, 1984].

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Philippine Sea Plate

34°-

142'E

Fig. 7. Tectonic setting of the 1923 Kanto earthquake illustrated on the 1885-1921 scismicity map in Fig. 6. Thick solid lines represent plate boundaries. Arrows indicate directions of relative plate motions at each boundary. The thick broken line shows the mechanical plate boundary between the Pacific plate and northeast Japan [ISfflBASHI, 1985c]. Thin solid and broken lines arc parts of depth-contours of the upper surfaces of subducted Pacific and Philippine Sea slabs, respectively, estimated by ISHIDA [1986] (with 20 km intervals). Hatched areas represent the source region of the 1923 Kanto, main-shock, major aftershocks sequence (as for K and Y, sec text). The star mark in the source region indicates Aburatsubo tide-gauge station. For three major earthquakes before 1923 the occurrence years are shown. IBT, Izu-Bonin trench; IZ, Izu Peninsula; JT, Japan

trench; SGT, Sagami trough; SRT, Suruga trough.

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Mechanisms of Seismic Quiescences

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Abstract

In the past decade there have been major advances in understanding the seismic cycle
in terms of the recognition of characteristic patterns of seismicity over the entire tectonic loading cycle. The most distinctive types of patterns are seismic quiescences, of which three types can be recognized: post-seismic quiescence, which occurs in the region of the rupture zone of an earthquake and persists for a substantial fraction of the recurrence time following the earthquake, intermediate-term quiescences, which appear over a similar region and persist for several years prior to large plate-rupturing earthquakes, and short-term quiescences, which are pronounced lulls in premonitory swarms that occur in the hypocentral region hours or days before an earthquake. Although the frequency with which intermediate term and short term quiescences preceed earthquakes is not known, and the statistical significance of some of the former has been challenged, there is a need, if this phenomena is to be considered a possibly real precursor, to consider physical mechanisms that may be responsible for them.
The characteristic features of these quiescences are reviewed, and possible
mechanisms for their cause are discussed. Post-seismic quiescence can be readily explained by any simple model of the tectonic loading cycle as due to the regional effect of the stress-drop of the previous principal earthquake. The other types of quiescence require significant modification to any such simple model. Of the possibilities considered, only dilatancy hardening seems viable in predicting the observed phenomena. Intermediate-term quiescences typically occur over a region several times the size of the rupture zone of the later earthquake and exhibit a relationship between the quiescence duration and size of the earthquake: they thus involve regional hardening and agree with the predictions of the dilatancy-diffusion theory. Short-term quiescences, on the other hand, are more likely explained by fault zone dilatancy hardening within the a small nucleation zone. Because seismicity is a locally relaxing process, seismicity should follow a behavior known in rock mechanics as the Kaiser effect, in which only a very slight increase in strength due to dilatancy hardening is required to cause quiescence. This is in contrast to other precursory phenomena predicted by dilatancy, which require large dilatant strains and complete dilatancy hardening.

Introduction

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Seismic quiescences are, aside from immediate foreshocks, the most well
documented and often reported phenomena that are precursory to earthquakes. This is not a positive indication that this phenomenon is more common than other precursory signals, since it probably reflects the fact that seismicity is the only earthquake-related process that is routinely monitored on a world-wide basis. Nonetheless, the frequent observation of seismic quiescences provides for it a credibility that demands serious consideration of possible mechanisms that could be responsible for producing this phenomenon and in turn to consider the consequences of this on other possible precursors.
There have been a number of reviews of seismic quiescences as part of the larger
subject of seismicity patterns (Rikitake, 1977; Kanamori, 1981; Reyners, 1981; Mogi, 1985; Wyss, 1985). The type of quiescence that is most often discussed in conjunction with earthquake prediction is one that may be classed as an intermediate-term precursor: it occurs over a relatively broad region for a time period of typically several years prior to the occurrence of the principal earthquake. Although this type of quiescence is the main subject of this paper, two other types of quiescences can also be recognized and merit discussion in this context. The first of these is post-seismic quiescence, which although not relevant to earthquake prediction, is almost universally observed and provides a readily understood mechanistic framework for understanding the general phenomenon. The second of these is a pronounced lull in seismicity that sometimes occurs between foreshocks and mainshock. This latter is therefore a short-term precursor, but except for scale it appears very similar to intermediate-term quiescence and its mechanism may also be similar.
The credibility of many reported cases of intermediate-term quiescences has been challenged because they have not been supported by rigorous tests of statistical significance (Reasonberg and Mathews, 1987) and in some cases may be artifacts of catalogue heterogeniety (Habermann, 1982,1987). It is also not clear how common this phenomenon is, because there have been no studies that have attempted a uniform census. These several objections have led the subject of seismic quiescences, like most
topics in earthquake prediction, to become controversial. Nonetheless, among workers in the field there seems to be a concensus that at least some subset of reported quiescences are real and precursory phenomena. It therefore seems legitimate to address the question as to possible mechanisms that could cause them.

Types of Quiescence

If we examine the entire loading cycle between successive ruptures on a specific
section of active fault or plate boundary a characteristic pattern of minor seismicity can be recognised. This "seismicity cycle", introduced by Fedotov (1968) and elaborated upon by Mogi (1977,1985), is illustrated in Figure 1, which is modified from Mogi (1985) with a few deletions and additions.
The "principal rupture" is followed by an aftershock sequence, A, which, though commonly concentrated near the ends of the rupture, decays hyperbolically in time into the post-seismic quiescence, Qi. Qi occupies an appreciable portion, typically 50-70%, of the recurrence period T, and usually extends over the entire region surrounding the

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rupture zone. This is followed by a general increase in "background" seismicity, B, over the whole region (Mogi, 1981, refers to this as the "active period" and considers these a type of foreshock, but we avoid this latter terminology as being somewhat confusing). This is commonly followed by an intermediate-term quiescence Q2, which typically extends over the entire zone surrounding the rupture and lasts for several years. If Q2 does not extend to the surrounding regions, where seismicity may be augmented, a "doughnut pattern",D, appears (this term is somewhat of a misnomer; the doughnut pattern seldom forms a complete ring: more often it is concentrated only at two ends of the rupture [Perez,1983]).
The succeeding principal rupture is preceeded by the immediate foreshocks, F,
which typically begin appearing about 10 days before the rupture (Jones and Molnar, 1979), and are usually concentrated close to its hypocentral region. If an appreciable number of foreshocks occur and are detected, a pronounced lull in this activity, (£3, is often observed for several hours to a few days prior to the mainshock.
The patterns shown in Figure 1 do not universally occur in a complete sequence and there is enough variation in individual patterns to make their recognition difficult in real time, yet they occur often enough to make such a synthesis possible, which suggests a fairly simple underlying cause. In order to discuss Qj_, Q^, and (£3 , the post-seismic, intermediate-term, and short-term quiescences in more detail, we next provide a few specific examples of each.
Post-seismic quiescences. Fedotov (1968) is usually credited with first pointing
out, with regard to large interplate earthquakes in the Kuriles and Japan, that prolonged periods of seismic quiescence follow the aftershock sequences of previous large earthquakes, and that it is only in the latter part of the loading cycle that activity picks up. In California, Tocher (1959), noted that seismic activity in the region of the 1906 California earthquake had been much greater in the 50 years prior to that earthquake than in the 50 years following it. Brune and Alien (1967) found that the rupture zone of the 1857 California earthquake was anomalously quiet at the microearthquake level, suggesting that it too was in a quiescent period.
Tocher's observation was made on the occasion of a study of a moderate sized earthquake that occurred on the San Francisco Peninsula in 1957, an earthquake that can now be recognized as marking the approximate beginning of a new active period in the region of the 1906 earthquake. The 1906-1956 quiescence is now a well defined post-seismic quiescence of regional extent, and serves as a good example here, as its properties have been described by Ellsworth et al. (1981) and Raleigh et al. (1982). Seismicity in the San Francisco Bay region is shown in 25-year intervals from 1855-1980 in Figure 2. The lull in activity during the period 1906-1956 in the region adjacent to the 1906 rupture of the San Andreas is quite apparent. Most of the activity during that time occurred south of the 1906 zone. The more recent buildup is exemplified most recently by the 1979 Coyote Lake (M=5.9) and 1984 Morgan Hill (M=6.2) earthquakes which ruptured adjoining sections of the Calaveras fault just opposite the southern part of the 1906 rupture. A characteristic feature of the recent active period as well as the pre-1906 active period is that they are marked by earthquakes on the Calaveras, Haywards, and other lesser faults to the east of San Francisco Bay rather than by enhanced

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activity on the San Andreas fault itself. The 1906 section of the San Andreas fault is still remarkably quiet, both for moderate sized earthquakes and microearthquakes (Bolt and Miller, 1975; Olson, 1986). Therefore, these active periods could be considered a kind of "doughnut" pattern. In the case of other active periods of the type labelled B in Fig. 1, which have been documented for subduction zone earthquakes, the resolution in not sufficient to determine if they are also of this "doughnut" type.

Intermediate-term Quiescences. Many examples of this type of quiescence have
been given by Kanamori (1981). We chose as our type example the case of the 1978 Oaxaca earthquake both because it is well documented and because it served as the basis for a successful prediction (Ohtake et al., 1977, 1981). The time history of this quiescence is shown in Figure 3 and its spatial development in Figure 4. In mid-1973 the quiescence began abruptly and simultaneously over a region 270 km in linear extent and an area of approximately 7x10 km . This area is much larger than the aftershock zone of the M=7.8 earthquake that followed in 1978, hence the quiescence is a regional phenomenon that is not restricted to the eventual rupture plane.This quiescence was terminated by a period of renewed activity that began 0.9 years before the mainshock, and which had been anticipated on the basis of seismicity patterns before other Mexican subduction zone earthquakes (Ohtake et al., 1977,1981). The final foreshocks occurred as a cluster beginning 1.8 days before the mainshock, followed by a short-term quiescence in the final 12 hours (McNally, 1981).
Ohtake (1980) made a survey of intermediate-term quiescences and found that, like
the Oaxaca example, they typically occur over a region larger than the aftershock zone of the earthquakes they preceed, by factors ranging from one to ten. He also found that such quiescences were followed by brief periods of renewed activity in about 35% of the cases. In addition, he found a relationship between the time interval from the appearance of the quiescence to the mainshock and the size of the mainshock (see Reyners, 1981, for a review of such relationships). Using data from Kanamori (1981), Mogi (1985) found a similar relationship, which is reproduced in Figure 5. Although the considerable scatter in this data makes this relationship of questionable utility for the purpose of making predictions, the data do support the contention that such a relationship exists, which provides strong constraints on the types of mechanisms that can be considered to cause quiescences. A similar relationship was found by Wyss and Habermann (1987), based on a different data set.

Short-term quiescences. As in the case of the Oaxaca example, earthquakes are
often immediately preceeded by a swarm of foreshocks, typically restricted to the hypocentral region, followed by a pronounced lull in activity. Perhaps the most well known case is that of the 1975 Haicheng earthquake (M=7.5, Raleigh et al. 1977). We refer to this as a premonitory swarm, noting a distinction to be made with a similar term used by Evison (1977) and Kanamori (1981) to describe enhanced activity that sometimes preceeds intermediate-term quiescences. In contrast, the premonitory swarm we discuss occurs a few days or hours before the mainshock, and the short-term quiescence that follows and defines it is of similar duration. Jones and Molnar (1979), in their general

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survey of foreshocks, found a lull in foreshock activity 5-10 days before mainshocks. Although this observation may be a detection of this phenomenon, generally the resolution of their study, one day, is too coarse to detect short-term quiescences, which often have durations of only a few hours.
Several examples of premonitory swarms followed by short-term quiescences are shown in Figure 6 for some earthquakes in the Izu Peninsula, Japan. The range in magnitude for these three mainshocks is considerable, being 7.0 for the 1978 Izu-Oshima-kinkai earthquake, 5.4 for the Kawazu and 3.6 for the Hokkawa shocks. In each case, the premonitory swarm consists of earthquakes about two magnitude units smaller than the mainshock, but the time scales for both swarm and quiescence clearly do not depend on the size of the mainshock. The swarm preceeding the Izu-Oshima-kinkai earthquake was tightly restricted to the hypocentral region of the mainshock (Tsumura et al., 1978), which is a characteristic of premonitory swarms before Izu earthquakes (Mogi,1985).

Mechanisms of Quiescences

In considering the significance of seismicity patterns, some general points regarding scale need to be made. First of all, the seismic moments of earthquakes in the seismicity pattern preceeding a major earthquakes are orders of magnitude smaller than that of the mainshock, hence they themselves play no significant role in affecting the loading process, since their moment release rate is insignificant. Rather, they can be considered a kind of stress gauge. Minor seismicity.is extremely sensitive to small changes in stress: the general experience of induced seismicity is that changes in load or pore pressure in the order of 1 MPa or smaller will bring about pronounced seismicity changes (e.g. Simpson, 1986). Although in cases of reservoir impoundment increases in seismicity have been most often noticed, in thrust environments the water load can produce a quiescence, as was the case of the Tarbella Reservoir in Pakistan (Jacob et al., 1979).
The other point regards the nonuniformity of the time scale in Fig. 1, and most other similar diagrams. The duration of the Qi» and Q^ quiescences have been greatly exaggerated there to make them visible. Recalling that the duration of a typical intermediate-term quiescence is 3-5 years for a large plate boundary rupturing earthquake, which might have a typical recurrence time of 100 years, the Q2 duration would only be 3-5% of T, and if plotted at a true scale it would occupy a very narrow space in Fig. 1. The duration of the short-term quiescence, being only a few hours or days, is even more exaggerated in the figure. Although it seems an obvious point, the true duration of the quiescences is very important in considering their mechanisms, and a superficial glance at a figure like Fig. 1 will be quite misleading in that respect.
The explanation of the origin of post-seismic quiescence is trivial, arising naturally
from the tectonic loading cycle in a heterogeneous lithosphere. However, it allows us to set up a basic model for the seismic cycle which must then be modified to account for the other quiescences. Consider that seismicity is controlled by a simple local fracture criterion:

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where Ty is a local frictional yield stress, defined not only on the principal rupture plane but at all points in the region, and Tt is the local tectonic stress. We then assume a loading function:

Tt =T0 + AT(t/T) (2)

where t is the time since the previous principal rupture, of stress-drop AT. To account for heterogeniety, iy and TQ can be considered to be random variables and we can describe the initial conditions at t = 0 with a probability density function /(i; 0), as shown in Figure 7a, where TC is the critical strength at which the principal ruptures occurs. If we then integrate over the loading function (2), the seismicity, /(e), is proportional to /(T; t), as in Fig. 7b. So in this simple model the post-seismic quiescence Qj is because the stress-drop of the previous principal earthquake reduced the stress regionally to a level below the local strength, and the resumption of activity in B occurs when it reaches some critical value, say Ta. This scheme is essentially the same as that described less explicitly by Mogi (1981).
If we consider the two-dimensional extent of the stress-drop, we can account for the "doughnut" pattern nature of B as pointed out in the case of the 1906 California earthquake. The stress-drop from the 1906 earthquake on the San Andreas fault also reduces the stress on the adjacent Haywards and Calaveras faults, causing them to become quiescent (Figure 8a). As the stress re-accumulates, these faults, which experienced a lesser stress-drop in 1906, will reach their critical stress and become active before the San Andreas fault re-activates (Figure 8b). A similar pattern is expected in other tectonic settings.
No model of this type, with a smooth strength distribution and monotonic loading,
will predict a quiescence of the Q2 type. There are three ways, in general, that one can modify this model to account for an intermediate-term quiescence of the type observed. The first is to invoke a regional reduction in it during the quiescent period, the second is to invoke a specialized strength distribution /(T; 0) with a bimodal form, and the third is to invoke some hardening mechanism that will produce a temporal increase in Ty. These three possibilities will be discussed shortly, but first we need to consider witn what sensitivity this system will respond to produce a quiescence. One might first think that Tt would have to drop below the level Ta, or the equivalent change in Ty must occur, before a quiescence would become evident. It is more physically reasonable, however, to make an "exhaustion" assumption whereby when the fracture criterion T = 0 is locally met there is a local stress drop that greatly increases T at that point so that it is no longer close to rupture. With this assumption the strength distribution /(T; t) becomes truncated to the left at T = 0, as in Figure 7c, so that an infinitessimal drop in it or rise in iy will result in a quiescence. This type of behavior is well known in experimental rock mechanics, where it is known as the Kaiser effect.. Thus a rock that has been previously loaded to a certain stress level, when reloaded will not produce any acoustic emission until it has been loaded to a higher stress than the previous one, and conversely any reduction in stress will cause acoustic

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emission to cease. This memory property is essential if we are to avoid having to invoke gross changes in stress or strength to explain quiescences.

Mechanisms involving tectonic stress changes. If we consider the simple loading model shown in Figure 9a, a quiescence will occur if there is a perturbation in the monotonic increase in tt such that it drops below its previous value for a period, as in Fig. 9b, resulting in the seismicity pattern shown in Fig. 9c, in which normal activity follows the quiescence leading to the principal event that occurs after a delay t^.
Such regional perturbation in the stress field are known, the most famous being the Palmdale uplift (Castle et al.,1976), and Thatcher (1983) has discussed their importance to the loading cycle (that discussion was prompted by a nonlinear stress change on the San Andreas fault reported by Savage et al.,1981 and later discounted by Savage et al.,1986). The mechanism of such phenomena is unknown, but there has been no suggestion that they are directly related to the earthquake loading cycle. Thus such phenomena are "demons" to be invoked to produce a desired affect. Such demons have been invoked before, to explain migration of earthquakes over large distances (one of these was suggested by the author, in Scholz,1977). Triggering demons of that type are very unspecialized as compared to any demon that might be called upon to produce a quiescence. The fact that intermediate-term quiescences seem to be common and have very clear spatial and temporal relationships with the principal rupture argues strongly that any mechanism must be intimately connected with the loading cycle: no such connection is known which would provide a reduction of tectonic stress at a specific point in the loading cycle. Furthermore the return to normal activity after the quiescence, predicted by this model, is not usually observed; when renewed activity occurs, as in the Oaxaca case, it is for a very brief period. This type of mechanism therefore does not seem satisfactory as a general explanation for intermediate-term quiescences.

Mechanisms involving specialized strength distributions. An alternative model could
be devised in which one retains the assumption of monotonically increasing tectonic stress but assumes a bimodal distribution in initial strengths, /(t; 0), as in Figure 9d. This model was proposed by Kanamori (1981). It assumes that there are two distinct populations of regions with different strengths, which fail in the period B and in the period of the principal rupture, the quiescence occurring because there are no regions of intermediate strength. This model suffers from being ad hoc: no explanation is given for why there should be two strength populations or why they should have such a pronounced 'hole 1 between them. Figure 9d is drawn approximately to scale, taking into account the relatively long duration of B and very short duration of Q2 relative to T, mentioned before. Thus the gap between the two strength populations would have to be very narrow to explain the relatively short duration of the quiescence. As an example, Kanamori (1981) tried to model the case of the 1963 Kuriles earthquake, which was preceeded by a 2 year quiescence. If the recurrence time for that earthquake is 100 years, say, then at most the separation between the two strength distributions could only be 2% of the total. This fine distinction is hard to rationalize on the basis of random heterogeniety and was not reproduced in Kanamori's models.

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A final point is that the relationship between quiescence duration and magnitude
(Fig.5) is incompatible with this kind of model, since the latter is based on an assumption on initial conditions. One could only predict it if one further assumed a relationship between the width of the 'strength gap' and the magnitude of the impending earthquake!

Mechanisms involving hardening. A final possibility is that there is some
mechanism whereby Ty increases temporarily, thus increasing i and producing a quiescence, as in Figures 9e and f. In order to account for the timing of the quiescence this effect must be coupled to the loading cycle and hence be a form of hardening. The only type of hardening likely to occur in the brittle part of the lithosphere is dilatancy hardening, where dilatancy reduces the pore pressure and thus increases the frictional strength of faults. Two types of dilatancy are possible: whole rock dilatancy in which a volume surrounding the ultimate rupture zone becomes dilatant, and fault zone dilatancy, in which the dilatancy is restricted to the materials in the fault zone. Present laboratory results indicate that the first will occur at stresses just below that necessary to cause frictional failure on a fault, but that the second can occur at much lower stresses, between 1/3 to 2/3 the frictional strength (Holcomb, 1978; Raleigh and Marone, 1986).
If we consider the Oaxaca case, and other cases summarized by Ohtake (1980) to be typical, they indicate that intermediate-term quiescence is not restricted to the rupture zone, but extends over a considerable region. We conclude that if hardening is the mechanism, then it is whole rock dilatancy hardening, and the mechanism for intermediate-term quiescences is the same as that given in the dilatancy-diffusion theory (Scholz et al., 1973). This mechanism is consistent the main aspects of the phenomena, including providing a mechanism for the duration-magnitude relation (Fig. 5). In some cases, however, these processes may vary in space, as in the case of the Kalapana earthquake, where quiescence was observed simultaneously with nearby enhanced activity (Wyss et al., 1981). As will be discussed more fully later, because of the Kaiser effect quiescence is much more easily triggered by dilatancy hardening than are other precursory phenomena, such as velocity anomalies, which are more commonly associated with this theory. This means that quiescence would be the most sensitive indicator of dilatancy hardening, and may occur without velocity anomalies or other dilatancy-related precursors.

Mechanism of short-term quiescences. The premonitory swarms and short-term quiescences described above are similar to the intermediate-term quiescence pattern except for their shorter duration and much smaller spatial extent. Rather than being spatially focussed on the entire rupture zone they are focussed on a small region near the hypocenter, which points to their being related to the nucleation process. As Rice (1979,1980) has pointed out, the stick-slip instability can be expected to be preceeded by accelerated stable deformation in a small region around the point of eventual unstable rupture initiation. If the fault zone material is dilatant, as Stuart (1974) and Rice (1979) have suggested, then local dilatancy hardening will occur within the nucleation patch, resulting in a temporary stablization (see also Rudnicki, 1986).
A sensible interpretation, therefore, is that a premonitory swarm signals the

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occurrence of nucleation, and the short-term quiescence that follows results from local dilatancy hardening of the nucleation patch. Since this is fault zone dilatancy, it is restricted not only to the small areal extent of the nucleation zone but also to a thin wafer, so that fluid diffusion paths are short leading to the brief duration of the phenomena. An interesting observation, based on this interpretation and the few examples that we have shown, is that the time scale of these phenomena do not appear to scale with magnitude (Fig. 6), which implies that the size of the nucleation zone, in at least one of its dimensions, does not scale with the size of the eventual earthquake.

Discussion

A prolonged post-seismic quiescence followed by an active period of background seismicity is a consequence of any simple model of the tectonic loading cycle. Intermediate-term preseismic quiescence cannot be explained without a significant modification of this model. The most likely explanation, consistent with the observed features of these quiescences is that they are caused by dilatancy hardening over a broad region, in the manner predicted by the dilatancy-diffusion theory.
When dilatancy-diffusion was first proposed, this type of quiescence had not yet
been recognized as a common earthquake precursor. Thus, although quiescence was predicted as a precursor by Scholz et al. (1973), most of the discussion of the theory at that time concentrated on other types of precursory effects, such as velocity anomalies and crustal uplift. So too the later unpopularity of this theory was a result of unsuccessful attempts to observe velocity anomalies (e.g. McEvilly and Johnson, 1974), or because some of the precursors marshalled to support the theory were later discredited (Mogi, 1985). The observations of intermediate-term quiescences reviewed here, however, are almost exactly as had been predicted by the theory, which requires a serious re-evaluation of if it as a viable possibility.
Consider the mechanism of dilatancy hardening, as shown in Figure 10 (after Brace
and Martin, 1968). The strength of rock at various strain rates is shown for two equilibrium conditions, one with a confining pressure pc = p j, saturated with pore pressure at pp =0, the other with p~= Pj-p2 and Pp=0. Compared with these is the strength of rock deformed 'drained at various rates with PQ=PI and Pp=p9. At low strain rates this followed the pc= P\~P2 curve> but above a critical strain rate, at A, the rate of dilatancy became faster than the rate of fluid inflow so that dilatancy hardening occurred and the strength increased above that curve. At a sufficiently high rate, at B, dilatancy hardening became total, and the strength reached the value for pc=pj, Pp=0. Because of the Kaiser effect, the slightest dilatancy hardening will result in seismic quiescence, which would thus begin at point A. The appearance of velocity and electrical resistivity anomalies, on the other hand, require that the medium become undersaturated, which will not occur until dilatancy hardening is complete, at point B. Similarly, anomalous crustal uplifts require large dilatant strains since cms of uplift must occur to be detected by geodetic means. Thus a slight amount of dilatancy could produce a seismic quiescence without producing any of the other dilatancy-related precursors, which all require gross dilatancy. Otherwise, the predictions concerning quiescence of Scholz et al. (1973) agree

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well with the observations of the intermediate-term quiescences: /, they occur within a large volume surrounding the rupture zone, //, they are often followed by a brief renewal of activity (expected if pore pressure returns to normal), and Hi, the duration of the anomaly is proportional the the size of the impending earthquake.

acknowledgement

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0 Finure 1. A schematic space-time diagram showing the types of seismicity patterns observed throughout the seismic cycle. Those elements discussed in the text are: A, aftershocks;
Q!» postseismic quiescence; B, background seismicity; Q2, intermediate-term
quiescence;
D, doughnut pattern, F, foreshocks, and 03, short-term quiescence.

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63 I 64 65 66 67 I 68 I 69 I 70 71 I 72 73 I 74 I 75 I 76 I 77 I 78

YEAR
Ms=7.7

Figure 3, Seismicity in the region of southern Mexico for two-year periods just before andduring the quiescence shown in Fig. 3 (from Ohtake et al., 1981). The a and B periods and Untake s notation for the quiescence and the period of re-activation

(a) JUN.1971 -MAY 1973 20°N

mb: -5-6-7

15

10'

(b) JUN. 1973-MAY 1975 20°N

105°W 100 95'

Figure 4. Seismicity in the gap of the 1978 Oaxaca, Mexico, earthquake, showing the intermediate-term quiescence (from Ohnake et al, 1981).

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A. Post-seismic

QUIESCENCE

B. Pre-seismic

ACTIVITY

Figure 8. Schematic stress profile perpendicular to the San Andreas fault for two periods, a) following stress relief by slip on the San Andreas, the adjacent faults also become quiescent,b) after a period of stress accumulation, the adjacent faults are reactivated but the San Andreas is still quiescent.

TIME

-*» (D

SEISMICITY

b 1 1
1 1
1 1

' ' 4

A?
TIME

f(T,o)

STRENGTH

TIME

f(e)

O

LU

TIME

Figure 9. Different modifications of the model of Fig. 7 that can account for intermediate-term quiescences. a,b) model and predicted seismicity based on the assumption of a perturbation in tectonic stress, c), model based on an initial bi-modal strength distribution, d,e) model and predicted seismicity based on a hardening assumption.

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COMPLETION OF DILATANCY HARDENING, VELOCITY,
RESISTIVITY ANOMALIES

ONSET OF DILATANCY
HARDENING,
QUIESCENCE
A

STRAIN RATE

Figure 10. Illustration of dilatancy hardening (after Brace and Martin, 1968). See text for discussion.

Summary of Session on Seismicity Patterns
by Robert L. Wesson

Ted Habermann began the session with a careful and thoughtful review of precursory seismic quiescence. He pointed out that the study of seismicity is particularly important for intermediate-term earthquake prediction because:
Seismicity data represent the largest data base available for such studies
It is possible to assemble numerous case histories of the seismicity preceding large earthquakes
It is possible to determine empirically the success and false alarm rates
The data are "remotely sensed," that is, measurements in the immediate epicentral area may not
be required

He indicated that the major difficulty in studying seismicity variations is the identification and correction for changes in the configuration, instrumentation, operation and data processing of seismic networks. As apparent, but not natural, seismicity changes, he distinguished "detection changes," "reporting changes," and "magnitude shifts," and illustrated techniques for testing seismicity catalogs based on identifying changes in the number of smaller earthquakes at times when the the number of larger earthquakes remains relatively constant. Examples of apparent seismicity changes caused by network changes include a detection increase in the world wide catalogs caused by the installation of the WWSSN in about 1964, and a decrease caused by the closure of the VELA arrays in 1967. An example of a seismicity variation argued to represent a real natural fluctuation was the quiescence preceding the 1983 Kaoiki, Hawaii, earthquake.

Kiyoo Mogi described an apparent downward migration of seismicity prior to the 1944 Tonankai and the 1952 Tokachi-oki earthquakes. In both these cases, the mainshock occurred after the the downward migration reached a limit (70 km for Tonankai, 400 km for Tokachi-oki). Professor Mogi noted that recent observations in the Tokai district of Japan fit a similar pattern, and suggested that downward migration of seismicity may be an intermediate-term precursor of great thrust-type earthquakes in subduction zones.

Karen McNally reported on seismicity patterns including clusters in California and quiescence, migration and pre-shocks in Mexico. The precursory clusters for California were typically formed of smaller earthquakes, less than magnitude 4, occurred within 2 to 3 times the source dimension of the subsequent mainshock, and preceded the mainshock by a few weeks to a few years. In Mexico, the precursory phenomena occur within the lateral dimension of an existing seismic gap. Probability gains were determined for subsequent mainshocks.

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Max Wyss described in detail examples of seismic quiescence. He defined seismic quiescence as a decrease of mean seismicity rate as compared with the preceding background rate in the same crustal volume. He reiterated that the principal problem in identifying seismicity rate changes is caused by the presence within seismicity catalogs of artificial rate changes caused by changes in network operations or processing. He indicated that precursory quiescence preceded 18 earthquakes with rate decreases ranging from 45 to 90%, for time periods ranging from 15 to 75 months.

Katsuhiko Ishibashi distinguished two types of intermediate-term earthquake precursors: physical precursors and tectonic precursors. Physical precursors are direct or indirect indications of the initiation of the a rupture process. A tectonic precursor is a manifestation of tectonic movement outside the crustal domain of the impending earthquake, but part of a chain of tectonic events leading to the earthquake. As examples of tectonic precursors, he cited water level and crustal deformation preceding the 1978 Izu-Oshima earthquake, seismicity patterns preceding the 1923 Kanto earthquake.

Chris Scholz distinguished three types of seismic quiescence within the context of the seismic cycle: post-seismic quiescence, intermediate-term quiescence, and short-term quiescence. He explained post-seismic quiescence as due to the regional effect of the stress drop from the preceding earthquake. He interpreted intermediate-term quiescence in terms of regional scale dilatancy hardening, and short-term quiescence in terms of fault zone dilatancy hardening.

Discussion at the session raised questions about the appropriate test for statistical significance of reported seismicity patterns (see also the subsequent session on Statistical Methodology). The concept of the tectonic precursor also generated considerable interest.

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Statistical methods for investigating quiescence and other temporal seismicity patterns

by

Mark V. Matthews
and
Paul A. Reasenberg

Abstract

We propose a statistical model and a technique for objective recognition of one
of the most commonly cited seismicity patterns: microearthquake quiescence.We use a Poisson process model for seismicity and define a process with quiescence as one with a particular kind of piece-wise constant intensity function. From this model, we derive a statistic for testing stationarity against a "quiescence" alternative. We discuss the asymptotic null distribution of this statistic. The final three sections of this paper are used to illustrate and discuss the restrictiveness of our model in particular and of the quiescence idea in general. We point out that there are many point processes which have neither constant nor quiescent rate functions. We emphasize the need for thor-ough testing for and description of nonuniformity in seismicity processes. We mention several commonly used distribution-free tests for uniformity, and we recommend some simple graphical methods which might be used to elucidate underlying patterns.

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§1. Introduction

Statistical approaches to seismology are both necessary and desirable. Because seismogenic processes are complex and largely unobservable, it is sometimes necessary to formulate statistical models based on observable data such as seismicity. In such models, simple stochastic components are used to represent "averaged" states of com-plicated physical ones. This approach is sensible because seismicity patterns ought to contain information about processes which generate them.

The most basic statistical problems in seismology concern the frequency distri-butions of earthquakes in time and/or space. In the earthquake prediction literature, there are numerous references to such well known spatio-temporal patterns as fore-shocks, doughnuts, seismic gaps, and quiescences (Kanamori, 1981). But some of the published work on seismicity patterns is marked by the failure of researchers to define clearly the patterns of interest or to specify techniques for their objective recognition. In the literature one finds many claims like "earthquake X was preceded by pattern P, so pattern P is one to look for if you want to predict future earthquakes." Such assertions may fail to establish the validity or reliability of pattern P as an earthquake predictor, however, if they neglect to consider how often P occurs without being closely followed by a mainshock, or how often mainshocks occur that are not preceded by P. Indeed the subjective methods and vague definitions sometimes employed in seismicity studies have made it difficult to address such issues as these.

This paper and its companion (Reasenberg and Matthews, 1986) consider tempo-ral seismicity patterns with particular attention to so-called microseismic quiescence. In Section 2 we propose a model for temporal seismicity distributions, give a for-mal statement of the "quiescence hypothesis" in terms of this model, and propose a statistical method for testing this hypothesis. In Section 3 we illustrate and discuss problems which arise in attempts to quantify or characterize nonuniformities in seis-micity distributions; in Section 4 we list several commonly used distribution-free tests for uniformity; and in Section 5 we emphasize the importance of graphical indicators and nonparametric estimators in elucidating underlying patterns in seismicity rate. Reasenberg and Matthews (this volume) employ the model and methods presented herein in a study of seismicity in Central California and Central Japan.

§2. A model and test for quiescence

For a given earthquake catalog, let S(t) denote the cumulative count of seismic events at time t for t in the observed time interval [0, T]. We think of the raw process

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5 as the sum of two components

S(t) = N(t) + D(t)

where D is the highly clustered "dependent events" process and N is the underlying "background" process in which we are interested. We use the algorithm described in Reasenberg (1985) to decluster 5 and isolate the process, TV, of present interest. We will denote the random occurrence time of the kth event in a sequence by 7* and X). will denote the kth interevent time.The random variables Tt and Xj are related by Xk = I* Tjfc_i. We always take the origin of the time axis as TI, the occurrence time of the first event, and we consider observation to have ceased at the time of the last recorded earthquake. Thus, we work on a time scale defined by TI = 0 and Tn = T. We model the point process N(t) as a (possibly) nonhomogeneous Poisson process with intensity function X(t). The nonhomogeneous Poisson process (NhPP) is an integer-valued process with unit jumps for which counts in nonoverlapping intervals are independent Poisson random variables. Specifically, if/ = [s,i], let C(I) = N(t) N(s) denote the number of events in /. Then C(I) is Poisson with mean

A(J)= /\(u)du.
Jl

For nonintersecting intervals /i,/2,C(/i) and C(/2) are independent Poisson random variables with means A(/i) and A(/2). See Ross(1983) for discussion of some properties of nonhomogeneous Poisson processes.
The quiescence hypothesis may be stated as a hypothesis about the shape of the intensity function in the NhPP model. Given N(t) on [0,T], let A be an unspecified subinterval of [0,T], and parametrize A as

(pB if t A
\(t) = (1)
v. B otherwise

where B (the "background" rate), p (the ratio of the "anomalous" rate to the back-ground) are positive constants, and A is some ("anomalous") subinterval of [0,T]. In terms of (1), the quiescence hypothesis is

HQ : There exists an interval A for which p < 1.

We wish to test HQ against the null hypothesis of stationarity, i.e., HQ : p 1. We propose to condition our test on the value of the random variable N(T) in (1).

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Say N(T) = n, and define time so that T = 1. A generic intensity function of type (1) is shown in Figure 1.
On the normalized time scale, the event times T\,..., Tn are distributed like the order statistics from a set of n independent random variables with common probability density function
_ , . /\(6 J . .

/(*) = ! (2)
/ \(u)du
o

For the subinterval [t 8, t] in the unit interval, define the random variable M(t, 8) as the number of Tj in the interval [t 8,t]. Then M(t,8) has a binomial distribution with parameters n and p(t, 8), where

t

= J f(u)du.
t-6

Under the null hypothesis, f(t) = 1 for all t [0,1], so p(t,8) = 8 for £, 8. Hence, the null expectation and variance of M(t,8) are E M(t, 8) n8 and Var(M(t,8)) = n8(l - 8). Define /? (*,£) by

- 6)

By the familiar asymptotic normal approximation to the binomial distribution, the null distribution of /3n (t,8) is approximately standard normal for sufficiently large n. Under HQ, on the other hand, if the quiescent interval is of length SQ ending at time £Q, then /3n (tQ,8Q) is asymptotically normal with mean

E{3n (tQ,8Q) = (p-

and variance

Var({3n (tQ, 6Q)) = p(l + 6(p - I))"2.
We expect, /3n (^Q,^Q) to be small, i.e., negative with large absolute value. It is ap-parent, in fact, that with probability arbitrarily close to one /3n (^Q, $Q) will be smaller than any fixed real number for sufficiently large n.
The prescription for testing the quiescence hypothesis HQ against uniformity is
to compute /3n (t,8) for all points (£, 8) in some index set X and to base a test on the value of the statistic

3n (t,8).

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The location of the quiescent interval is estimated by the interval at which /? is minimized

= mn

We remark that this is, asymptotically, the maximum likelihood estimate for the anomalous interval A and that our test is asymptotically equivalent to the gener-alized likelihood ratio test for HQ versus HQ. The maximum likelihood estimate for the rate ratio, /?, is
* m°

(n ma)£0 '

where Sa = SQ and ma = m(tfio).
To test the quiescence hypothesis we need to know the null distribution of the statistic 5~. Let W(t) be a standard Brownian bridge process on [0,1]. (The reader unfamiliar with the definition of this stochastic process may consult Ross (1983), pp. 187-189.) Denoting the index set of intervals by In to remind us that it may depend on the sample size, n, we find the following result easy to deduce from standard weak convergence arguments (see, e.g., Billingsley, 1968). For a suitably chosen sequence Xn of index sets, distribution of S~ is asymptotically approximated by the distribution of

. W(t) - W(t - 6) W = mm '
- 6)

This distribution depends, of course, on / . We generally take the index set to be the square lattice of size SQ over the triangle 0 < 8 < t < 1. We have obtained approx-imations to these distributions by simulating Brownian bridge processes to generate samples of 10, 000 values of W~ over various index sets. Table 5 of Reasenberg and Matthews (this volume) lists some quantiles of the simulated distribution of W~ over the lattice with SQ = 1/300.

§3. Some examples
We exemplify application of the test described in the previous section on three synthetic data sets shown in Figures 2, 3, and 4. The density functions from which these samples were drawn are listed below along with the results testing for quiescence.

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Example 1:

n = 575

/(*) = 1
5~ = -2.32 on [.01,.93]
Example 2:
n = 575

.4, if t E [0.5,0.75]

1.2, otherwise
S~ = -8.45 on [.50, .74]
Example 3:
n = 575

(i + (t-i)2)

5~ = -4.85 on [.30, .72]
In Example 1, no significant anomaly is found in a sample that is truly uniform.
In Example 2, our test finds significant quiescence and essentially pinpoints the rate decrease in the third quarter of the observed time interval. Given the estimated qui-escent interval in Example 2, the estimated rate ratio is p = .316, which is very close to the actual value of 1/3.

In Example 3, we again find a significant rate anomaly. If this sample represented actual earthquakes, we might assert that our numerical finding indicates a significant quiescent interval. This finding is misleading, however. The value of the density, or intensity, function in Example 3 is indeed low on [.3, .72], but this interval represents a dip in a smoothly varying quadratic intensity function. It is not an anomalous departure from some well-defined long-term uniformity.
These three examples only begin to illustrate a basic fact. Namely, there are many possible seismicity distributions of which relatively few are included in any rea-sonable parametric family. Students of spatial and temporal earthquake distributions often specify and search for some "interesting" pattern which we'll generically call P. Usually the pattern is described by just a few parameters and the question is, "Does this set of earthquakes exhibit P or is the distribution uniform?" This procedure is commendable when it leads to useful detection, measurement, and description of important distributional features. But it can mislead. Care must be taken that, in the pursuit of rigorous, concrete results, we are not fooled by asking overly simplistic questions.

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If we partition the universe of possible intensity functions in terms of a pattern, P, then we have three sets: uniform distributions, distributions with P, and distributions which are neither uniform nor P. For any simply parametrized pattern - a quiescent interval, for instance - the class of P distributions represents a very small subset of all possible distributions. The vast majority of distributions are, like the one in Example 3, neither uniform nor P. The procedure of writing down a test statistic for P, calculating this statistic for a given data set, and reporting whether or not the statistic is significant is generally unlikely to achieve our goal of recognizing and accurately describing whatever "signal" may exist in a seismicity distribution. We must rely on more general methods, often graphical, to help us to describe qualitatively the underlying distribution and to interpret the quantitative results of hypothesis tests.

§4. Tests for uniformity
A basic question, always of interest when studying seismicity, is, "Do these data come from a uniform distribution?" When testing uniformity against a particular (class of) alternative(s), we ought to use a test designed to detect whatever specific departure from uniformity we might expect. In the common situation where we wish to test against general nonuniformity, several omnibus test statistics are available. Letting Fn denote the empirical distribution function, then we have the Kolmogorov-Smirnov statistic

Dn = \/n max |Fn (x) xl, x (0,l)
the Cramer-von Mises statistic
1
"2 = /(JUS) - X)2 dx,
0

and the Anderson-Darling statistic

1

f r F (r} rr i 2 TT7-2 Wr I * f*V*t/ / *' j

= n I . = ax.

J U/x(l-x)J

For computational formulas and discussion of the use of these statistics, see Cox and Lewis (1966). We recommend calculation of each of these statistics, possibly even in conjunction with other tests. If several tests for uniformity yield the same result, then we have the assurance of well-corroborated evidence for uniformity or nonuniformity. If some tests accept uniformity while others reject it, then we may look at the power of each test relative to various alternatives to determine what kind of nonuniformity

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might be more easily detected by the former set of tests than by the latter. Reasenberg and Matthews (this volume) have used all three of the aforementioned test statistics as well as the statistic

Sn = max \(3n (t,6)\.

We remark that the uniform data in Example 1 passed all four tests of uniformity, while the data from Examples 2 and 3 failed all four at the .01 level.

§5. Graphical indicators and noiiparametric estimators
of the intensity function
In the present context, there is a natural and useful graphical supplement to a numerical test stemming directly from the test itself. We must compute (3n (t,8) for many (t,6) values in order to determine the value of S~. Rather than simply printing out the minimum value of /? , we may examine the entire process over the domain J by plotting its contours, for instance. This way we are able to see where /? is low, where it is high, and how it behaves in intermediate regions. Figures 5-7 show contour plots of 0n for the samples of Examples 1-3. These particular plots were drawn with contours spaced every 3.84 units about zero, this spacing being the same as the 90% critical point in the distribution of S~ given in of Reasenberg and Matthews (this volume). Thus a closed contour (other than at /? = 0) encloses a region in which (3n is in sojourn below (or above) a level which it would cross in only one case in ten if applied to uniform data. Interpretation of these contour plots generally takes a bit of thought, but the basic patterns relative to the average rate are quite clear in our three examples. In Figure 5 we see only the zero level contour and a scattering of local extrema which are not too large in absolute value. There is little evidence of any pattern in this plot. Figures 6 and 7 show deep "valleys" of /? in regions where the underlying rate functions are low. They also show that there are significant high values of /? . In fact, if we let S+ = max^^gj (3n (t, £), then S+ = 5.086 on the interval [0,.45] for Example 2. For Example 3, S+ = 3.37 on [.72,1.0]. We see that, as we expect, these processes spend little time in the vicinity of zero when there are large rate fluctuations. This points to a difficulty of defining a "background" rate in many processes where there are apparent intervals of "high" and "low" rate. The symmetry between these intervals makes it difficult to assert empirically that one time period represents the background and that another is an anomalous departure. To analyze further these plots one could compute the mean function of /? and observe that the contours in these plots roughly follow the contours in their underlying mean functions. A useful exercise for a user of plots such as these would be to think about what sorts

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of patterns to expect from various hypothetical intensity functions and then to look for these patterns in plots produced from real data.
Certain graphical methods, examination of space-time plots or cumulative count functions, for instance, have found wide use in seismicity studies. Some of these methods are usually quite poor at providing meaningful insight into what is "really going on" in a local seismicity distribution. When looking for "significant" patterns in a seismicity distribution, the statistical issue is whether some apparent feature in the data is or is not reasonably likely to have arisen by chance given the perceived variability in the data. Visual estimates of stochastic variability are unreliable since they depend on such arbitrary parameters as the size of the plot and the scale relative to the sample size. To elucidate and describe an underlying signal which is "smooth", in some sense, we need to smooth our raw data. Ideally, we smooth out distracting random fluctuations to obtain a clear picture of the estimated underlying signal. In the present context, we are trying to estimate the probability density function specified in (2). Suppose our assumptions about the form of the density function fall short of postulating that / belongs to some parametric family indexed by a low dimensional parameter. Then we have an estimation problem that is nonparametric in the classical sense that the estimand lies in an infinite dimensional space.
Statisticians have recently devoted considerable attention to the problem of non-parametric density estimation. Book-length treatments of the subject have been pro-duced by Tapia and Thompson (1978) and Silverman (1986). Many methods have been proposed for this problem. As an alternative to the familiar histogram method of density estimation, we have used the so-called "kernel method." Kernel estimates have certain theoretical superiorities to histograms. Also, unlike histograms, they are gen-erally smooth, continuous, and well-suited to visual inspection. (The interested reader may consult Chapter 2 of Silverman's book for a survey of density estimation tech-niques and Chapter 3 for a detailed discussion of the kernel method for one-dimensional data.)
A kernel estimate of the density / at some point x based on a sample t\,..., tn from / is given by

In this formulation, the kernel, K, is generally taken to be a probability density function symmteric about zero; the "bandwidth", /i, is a positive parameter which controls the

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amount of smoothing. One may intuitively view this density estimate as the result of "smearing out" the point masses at the sample points t\ , . . . , tn by convolution with the smooth function K.

To produce a kernel density estimate, one must choose K and h in (3). We have always taken our kernel to be the standard Gaussian density

K(u) = <t>(u) = (27r)- 1 /2 e-"2 /2.

Aside from the fact that the density estimate /^ inherits the smoothness properties of the kernel, the choice of K is not critical. The choice of h, on the other hand, is quite important. The bandwidth in a kernel estimate controls the amount of smoothing in much the same way as does the binwidth in a histogram. By varying the bandwidth from small values to large values, we can go from an extremely rough estimate, nearly reproducing the raw data, to a very smooth, flat estimate. This is exemplified in Figure 8, where we see three estimates of the density function from Example 2. In this case, where we happen to know what the true density looks like, it is easy to draw conclusions. The first estimate, with h .0048, is much too rough. The third, with h .48, is too smooth since the dip between 1/2 and 3/4 is not apparent. The second estimate, with h .048, is a fairly good likeness of /. (In a case such as this, where there are actual jump discontinuities in the underlying density, a well chosen histogram would actually be preferred to a smooth kernel estimate.)
There is no way to escape completely the subjectivity involved in selection of the bandwidth parameter. Several methods for automatic bandwidth selection have been proposed, however. We have used one such method, which, along with others, is described by Silverman (1986). This method, known as "least-squares cross validation"

*\

(LSCV), is based on the following argument. Suppose that fh estimates /. Measuring distance in £2 we have

= /(/*-/)*

Let

Then chossing the h which minimizes R will also minimize err since R(h) and err(h) differ by a constant (/ /2). If we evaluate / at a set of points yi, . . . , y then we may

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approximate the first term in (4) by

m

Now define /_, to be the density estimate obtained by omitting the ith sample point, i.e.,

(n -

(This 'leave one point out" idea is a form of a general technique which statisticians call cross validation .) Also define

(6)

It may be shown that E MQ(/I) = E R(h), so MO(/I) //2 is an unbiased estimator of Eerr(h). Hence, a reasonable strategy for choosing the bandwidth, assuming that the minimizer of E MQ is close to the minimizer of MO, is to pick h to minimize MQ. Various algorithms are usable for locating the bandwidth which minimizes Mo(h) for a particular sample. We have used an algorithm suggested by Silverman (1986) which makes use of the convolution form of kernel density estimates. One can obtain quick and easy approximations to MQ(/I) by writing down and rearranging the Fourier transform of the right hand side of (6). The middle picture in Figure 8 shows the density estimate from Example 2 with h = hcv = -0483. The density estimates in Figure 9 and 10 are for the data of Examples 1 and 3, respectively, with respective bandwidths .0548 and .0436 also chosen by cross validation.
The graphical and nonparametric methods discussed in this section may appear
to contradict the spirit of previous sections where the emphasis was on classical, ob-jective testing of statistical hypotheses. The parametric hypothesis testing approach is certainly more convenient and more powerful than nonparametric methods when the hypotheses being tested are "correct" to a good approximation. But when we aren't so sure what are the right questions to ask, then we must analyze our data with thought and care. There is more to be learned by subjective, exploratory analysis than from the "objective" report that some "significant" pattern is present. One must learn to ask the right questions in order to find the right answers.

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B

pB

Background" Rate

Anomalous" Rate

0 T-6

Figure 1. A typical intensity function for a nonhomogeneous Poisson process with quiescence. Events are produced at rate B for all observed time except in the hatched interval A. In this quiescent period, the rate is lower by a factor p.

CO
c
CD

o

CO
c
CD
O

.5

Time

Figure 8. Kernel density estimates for data from Example 2. The bandwidths are, from top to bottom, .0048, .048, and .48. The second of these was selected by least squares cross validation as the optimal width for this sample. 613

C0
c
Q>
Q

. 5

Time

Figure 9. Kernel density estimate for data from Example 1 with LSCV bandwidth
h = .0548.

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CO
c
CD

O

. 5

Time

Figure 10. Kernel density estimate for data from Example 3 with LSCV bandwidth

h = .0436.

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Precursory Seismic Quiescence:
A Preliminary Assessment of the Hypothesis

by

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ABSTRACT

Numerous cases of precursory seismic quiescence have been reported in recent years. Some investigators have interpreted these observations as evidence that seismic quiescence is a somewhat reliable precursor to moderate or large earthquakes. However, because failures of the pattern to predict earthquakes are generally not reported, and because numerous earthquakes are not preceded by quiescence, the validity and reliability of the quiescence precursor have not been established.

We have analyzed the seismicity rate prior to, and in the source region
of, 37 shallow earthquakes (M 5.3-7.0) in central California and Japan for patterns of rate fluctuation, especially precursory quiescence. Nonuniformity in rate for these pre-mainshock sequences was relatively high, and numerous intervals with significant (p < 0.10) extrema in rate were observed in some of the sequences. In other sequences, however, the rate remained within normal limits up to the time of the mainshock. Overall, in terms of an observational basis for intermediate-term earthquake prediction, no evidence was found in the cases studied for a systematic, widespread, or reliable pattern of quiescence prior to the mainshocks.

In earthquake sequences comprising full seismic cycles for 5 sets of
repeat earthquakes on the San Andreas fault near Bear Valley, Ca., the seismicity rates were found to be uniform. A composite of the estimated rate fluctuations for the sequences, normalized to the length of the seismic cycle, reveals a weak pattern of a low rate in the first third of the cycle, and a high rate in the last few months. While these observations are qualitative, they may represent weak expressions of physical processes occurring in the source region over the seismic cycle.

Re-examination of seismicity rate fluctuations in volumes along the
creeping section of the San Andreas fault specified by Wyss and Burford (1985) qualitatively confirm the existence of low-rate intervals in volumes 361, 386, 382, 372, and 401. However, only the quiescence in volume 386 was found by the present study to be statistically significant.

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INTRODUCTION

Observations of seismic quiescence have been reported as precursors to
many moderate and large earthquakes. Review papers on seismicity patterns such as Kanamori (1981), von Seggern et al. (1981), and Rikitake (1979) report numerous convincing cases of precursory quiescence. These observations may suggest at face value that seismic quiescence is an established, reliable precursor. Unfortunately, diversity among the cases reported makes it impossible to draw such a conclusion with certainty. Important differences among reported cases are found in the definition of quiescence and the method used to detect it, spatial volumes of quiescence and criteria for their association with the mainshock, magnitude level considered, and treatment of aftershocks and man-made rate changes in the catalog. Despite these differences, however, many cases of quiescence are individually convincing. When considered collectively they make a plausible prima facie case that some form of seismic quiescence may occur before some earthquakes.

To be useful in the prediction of subsequent mainshocks, a seismicity pattern must be temporally stationary (i.e., the pattern must repeat over time in the same place) (von Seggern, et al., 1981). This criterion can be met, however, only for repeated mainshocks with relatively short interevent times, (limited by the extent of the existing seismic catalogs), for which a sufficiently numerous sequence of pre-mainshock seismicity is available. Unfortunately, very few cases of this kind exist. In view of this practical limitation, demonstration of spatial stationarity of the proposed pattern (i.e., demonstration that the pattern occurs in many places) is usually considered sufficient to establish it's validity.

Von Seggern et al. (1981), alluding to the collected body of reported precursory seismicity patterns, termed the quiescence precursor as "somewhat well-established" because such reports were numerous and because the reported precursor seemed to scale linearly in time with magnitude of the impending earthquake. They preceded to test objectively the real lability of seismic precursors, including quiescence, foreshocks, migration, clustering and b-value, in the time range of a few hundred days, by searching for a common underlying pattern that could be used for intermediate-term or short-term prediction. Based on this study, which included over 500 mainshocks (M 5.8-7.0) drawn from the LASA and NORSAR arrays and the NEIS catalog for a variety of tectonic regions, they concluded that evidence is lacking for widespread (i.e., spatially stationary), nonrandom behavior of seismicity parameters just prior to mainshocks. At the 90 percent confidence level, less than 30 percent of the mainshock regions were found to exhibit precursory seismicity changes. A limitation of that study was the lack of data at low magnitude level below the mainshock magnitudes. So, the possibility that clear nonrandom patterens in seismicity may exist at lower magnitudes than those considered could not be ruled out. In the five years since the publication of that study numerous new observations of precursory patterns have been reported, and a generally optimistic view of the utility of seismic quiescence patterns has emerged. Today, this view is held by many investigators, and seismic quiescence appears to be among the most seriously considered candidates for an intermediate term earthquake precursor (Kanamori, 1981; Cao and Aki, this volume; Habermann, this volume). Physical theory has been proposed to explain its presence (Scholz, this volume; Cao and Aki, this volume).

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The quiescence hypothesis states that decreases in the rate of seismicity
in or surrounding earthquake source regions precede the mainshocks in a systematic, non-random way, such that an observation of such a rate decrease provides information about the time and location (and possible magnitude) of a future earthquake. The hypothesis has been applied to several postdiction studies (in which the mainshock had already happened), and positive results have been taken as partial validation of the hypothesis. Recently two predictions (one successful) of moderate size earthquakes were based largely on observations of microseismic quiescence in central California (Wyss and Burford, 1985). However, notwithstanding numerous convincing reports of precursory quiescence, we suggest that the validity of the quiescence hypothesis has not been established in the literature. Only a fraction of all earthquakes for which data are available have reportedly been preceded by demonstrable quiescence. These cases were, presumably, selected by the investigators from among all which they may have considered. While we acknowledge the utility of reporting selected cases of anomalous observations consistent with a particular hypothesis (such as the quiescence hypothesis), we assert that interpretation of such a selected set to establish the validity of the hypothesis introduces bias because failures are generally not reported. One cannot necessarily infer from the body of published reports that precursory seismic quiescence is a spatially stationary pattern, (i.e., one generally or systematically observed). An alternate hypothesis - that random fluctuations in seismicity account for the reported cases of precursory quiescence, and selective reporting creates the impression that these cases represent a systematic effect - cannot be rejected by the body of results reported to date.

To put these hypotheses into a context in which they may be more fairly evaluated, we have attempted to provide a new baseline study in which a broad suite of observations of seismicity rate fluctuation are considered. We examined 47 cases drawn from 3 earthquake catalogs for Central California and Japan. We have attempted to treat the data uniformly in the removal of aftershocks, correction for known man-made rate changes, selection of areas studied, and statistical method used to identify the periods of anomalous rate. We report all the results uniformly without preference for "successful" cases. For each of the selected study areas, we adopt a null model for the declustered and magnitude-corrected sequences consisting of a homogeneous Poisson process with intensity estimated from the data. The alternative hypothesis is that a non-homogeneous Poisson process (i.e., one with intensity varying in time) underlies the observed sequences, and that the variations contain a nonrandom pattern of quiescence before mainshocks. (The terms "intensity", "rate" and "density" are used interchangeably in this paper to refer to the intensity parameter of the underlying process). Application of statistical methods described in Matthews and Reasenberg (this volume) allows us to measure the nonuniformity of each of the selected sequences, and to detect significant departures of the estimated intensity from the homogeneous model. The occurrences of the anomalous intervals are then compared with the time of a subsequent mainshock (for those sequences followed by a significant mainshock) in an effort to recognize a precursory pattern in the observed intensity fluctuations.

The scientific questions addressed in this study are: 1) How does the estimated intensity of microseismicity fluctuate in the regions immediately surrounding moderate to strong earthquakes, in the intermediate term preceding

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the earthquakes? 2) In particular, is there a widespread or common pattern of precursory quiescence over the intermediate term that can be recognized, and thus used to predict the oncoming earthquakes? 3) What characteristic pattern, if any, exists in the fluctuations of intensity in the source region, over an entire seismic cycle? 4) How does the intensity of seismicity fluctuate in selected volumes of the seismically active, creeping portion of the San Andreas fault!

DATA

Data for this study were drawn from catalogs for central California
(CALNET), central Japan, (NRCDP), and Japan, (JMA), (Table 1). Data were selected from the CALNET catalog for the period 1974-1986; the period from 1969 and 1973 was not used because preliminary analysis indicated that the sensitivity in parts of the network prior to 1974 was not stationary, owing to the growth of the network in that period. Data were selected from the portions of the CALNET catalog in which the magnitude level of complete detection is 1.5 (Reasenberg, 1985). The NRCDP network, while only in existence since 1979, is similar to CALNET in terms of both its sensitivity and location in a region of intense shallow microearthquake activity (Okada, 1984; Ishida, 1984). Data were selected from the central portion of this network, where the magnitude level for complete detection is 2.0 or lower (Matsumura, 1984). Data from the JMA catalog were selected for the period 1926-1984. Completeness levels of magnitude for this catalog were estimated by Evison (1981) to be 5.5 for the period 1926-1979 and 5.0 after 1979. However, data were used in this study from the central (on or near land) portions of the catalog, where the completeness level is presumed to be lower, with magnitudes 3.0 and greater.

The study consisted of four parts. In each part we have attempted to
focus on one aspect of seismicity rate fluctuations, although it is apparent that these aspects are interrelated.

1. Pre-mainshock seismicity.

For each catalog, a set of earthquakes was selected consisting of the largest events in the catalog for which a lengthy and populous set of pre-mainshock seismicity data were available. For each of the mainshocks we examined the sequence of seismicity within a small region centered about its epicenter, beginning at the earliest time for which the catalog sensitivity is considered stationary, and ending at the time of the mainshock. To study the fluctuations in the pre-mainshock seismicity rate in each of the source regions we applied the methods developed for this problem by Matthews and Reasenberg (this volume).

The concept that a sub-interval of a seismicity sequence is in a state of quiescence involves three parameters: two parameters to specify the interval and one to describe the degree of quiescence. The fact that the intervals have a start time and a duration requires that the statistic used to detect quiescence be applied over their entire two-dimensional domain. Statistics defined on a one-dimensional domain, such as AS(t), LTA and others (Habermann, 1981a, b, 1982, 1983; Habermann and Wyss, 1984; Matthews and Reasenberg, 1987) test only a subset of the possible intervals and, as a result, may fail to

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identify significant intervals (because most intervals never get tested). The beta statistic (Matthews and Reasenberg, this volume), on the other hand, is defined for two-dimensional intervals (t - 5, t) for all possible values of t (interval end time) and 6 (interval duration). For a specified interval in a sequence of earthquakes, s(t, a) measures the difference in seismicity rate between the interval and its complement (the rest of the sequence). We calculated beta for each pre-mainshock sequence over all possible subintervals corresponding to grid points on the t-6 plane with spacing of 30 days. The distribution of the extrema of beta for a homogeneous process was determined empirically. Intervals with scores of beta outside the 0.10 critical point were identified and termed anomalous. Patterns of occurrence of these anomalous intervals, as a function of time prior to the mainshock provide a uniform basis for comparison of the precursory fluctuations in intensity among all the pre-mainshock sequences studied. The patterns of intervals are visually compared and searched for a common pattern. For the California data, 8 mainshocks (M 5.3-6.7) were used (Table 1). For each mainshock, seismicity between 0 and 20 km depth was selected from a circular region centered on the epicenter, with area approximately equal to the area of the mainshock's aftershock distribution. Radii of the areas of epicentral selection ranged from 2.0 km for M 5.3 events to 15 km for the M 6.7 event at Coalinga (1983). To correct for a suspected man-made systematic change in magnitude determinations made for events in the CALNET catalog (Ellsworth, private communication), reported magnitudes were increased 0.1 units for all events after 770428. Aftershocks were removed with the algorithm described in Reasenberg (1985). For the NRCDP catalog, 10 mainshocks (M 5.3-6.9) were used (Table 1). Seismicity between 0 and 60 km depth was selected from circular regions (radius 10-20 km) centered on the mainshocks epicenters. Aftershocks were removed with the same algorithm. For the JMA catalog, 15 mainshocks (M 6.0-7.0) between 0 and 60 km depth were selected (Table 1). Seismicity was selected for the same depth range with epicentral distance from the mainshock of 25 km or less. Aftershock removal for this catalog is believed to be incomplete.

2. Complete seismic cycles at Bear Valley, Ca.

A unique opportunity to observe directly the structure of seismicity in
the source region over a full seismic cycle is provided by the identification of repeat earthquakes on the San Andreas fault near Bear Valley, Ca. Sequences of seismicity occurring within spherical volumes surrounding 5 sets of well-located earthquake "doublets" (M 4.5-5.1) were studied (Table 2, Figure 1). The two events in each doublet closely agree in hypocentral location, display a high degree of waveform similarity and are of similar magnitude. Ellsworth et al. (this volume) identified these doublets and concluded that they probably are repeat events having a common source. The intervals between the events in each pair, which range from 8 to 14 years, are included entirely within the period of operation of the CALNET in the Bear Valley region. Ellsworth (1975) estimated the magnitude level of completeness for this region to be 1.0 during the early 1970's. We have chosen a conservative cut-off level of M ^ 1.5 for this part of the investigation. Seismicity was selected for the intervals between the events in each pair within spherical volumes of 2 km and 3 km radius surrounding the centroid of the hypocenters for the pair. The resulting sequences represent full cycles of seismicity in the source region, complete over a range of about 3 magnitude units. Application of the method of kernel density estimation (Matthews and

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Reasenberg, this volume), allows comparison of the average intensity fluctuations in the five sequences on a common timescale normalized to the length of the cycle.

3. Measurements of nonuniformity in intensity.

For the pre-mainshock sequences, the Bear Valley repeat event sequences,
and the sequences selected from the creeping section of the San Andreas fault (described below), we apply several well-known statistical tests of uniformity. In particular, we calculate Kolmogorov-Smirnov, Cramer-von Mises, and Anderson-Darling statistics (Matthews and Reasenberg, this volume; Cox and Lewis, 1966). Though none of these tests is superior to any of the others for detecting general alternatives to uniformity, it has been found that the Anderson-Darling statistic is more sensitive to departures from uniformity near the endpoints 0 and 1, than the Kolmogorov-Smirnov and Cramir-von Mises statistics, which are better for detecting departures near the middle of the unit interval. We also use the statistic S = max a(t, a) as a test statistic for uniformity. This statistic is designed to detect an interval of anomalous rate surrounded by intervals of background rate. Taken collectively, these four tests provide a good indication of the nature and degree of nonuniformity in a particular data set. If a sequence passes or fails all four tests at some level, then we are more certain that it is or is not uniform than we would be based on any individual test. On the other hand, a sequence passing some of the tests and failing others indicates nonuniformity of the type best detected by the failed tests.

4. Creeping portion of the San Andreas fault.

Fluctuations in seismic intensity within selected volumes along the creeping section of the San Andreas fault were observed with the beta statistic. This area was chosen for study in part because of its relatively intense microseismicity and its lack of large earthquakes. However, our original interest in these particular volumes was raised by Wyss and Burford (1985), who based two earthquake predictions on their observations of seismic quiescence there. Matthews and Reasenberg (JGR, in press) critisized the methodology that lead to these predictions on the grounds that the significance of observed low rate intervals was overestimated. Here we re-examine the seismicity in the same volumes with the methods described in Matthews and Reasenberg (this volume). To correct for possible man-made fluctuations in intensity introduced by suspected systematic changes in the determination of magnitudes in the Calnet catalog, the set of magnitude corrections used by Wyss and Burford (1985) were applied to these data. Aftershocks were removed with the algorithm described in Reasenberg (1985).

RESULTS

1. Pre-mainshock seismicity.

The results in this section are obtained for each seismicity sequence with
a method illustrated in Figure 2. In this figure four representations of the fluctuations of rate in a particular sequence of seismicity are shown on a common timescale. This 10.5-year sequence is for the 10 km-radius circular area surrounding the epicenter of the 1984 (M 6.2) Morgan Hill, Ca.,

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earthquake and ends at the time of the 1984 mainshock. The principal result obtained in this section relies on the identification of intervals of seismicity within each sequence with significantly high or low average rate. To find these intervals, e(t, «) is calculated for all possible intervals with ending time, t, and duration, 6, on a 2-dimensional grid with spacing 30 days by 30 days. The resulting 8385 values of beta are represented in Figure 2a by both contour lines and printed values corresponding to relative extrema. The contour levels are zero and multiples of ±_ 3.84, the levels corresponding to the one-sided 0.1 critical points. In Figure 2a only the (insignificant) zero-contour is shown because no value of beta exceeded the critical points in this case. In addition to the contours, relative extrema in e(t, 6) are shown. For example, the circled value of 2.6 shown in Figure 2a corresponds to a relative maximum in B at (1979.5, 1.9), corresponding to the 1.9 year interval ending 1979.5. This interval is indicated in Figures 2b and 2d, where the high rate of seismicity is apparent. For each sequence analysed, all intervals with rates outside the critical points are identified in this way. Figure 2c shows the familiar plot of cumulative number of earthquakes as a function of time, for the sequence. Figure 2d shows three estimates of seismicity rate for the sequence. The broken line represents the Poisson (constant) rate for the entire sequence. The fluctuating lines are time-varying rate estimates obtained by the kernel density estimation method (Matthews and Reasenberg, this volume). These estimates, obtained by a smoothing procedure utilizing averaging kernels with widths shown in the lower left corner of Figure 2d, allow easy visual interpretation of the rate fluctuation underlying the sequence. In this example, the 1.9 year interval of relatively high seismicity rate identified with beta, while not statistically significant, is easily seen as a prominant rate fluctuation. The smoother of the two rate estimates shown corresponds to the (longer) optimal kernal width determined by least squares cross validation (Matthews and Reasenberg, this volume). The less smooth estimate corresponds to a kernel width 0.5 times optimal.

This method for identifying significant rate fluctuations was applied uniformely to all the pre-mainshock seismicity sequences. The results are shown, for selected cased, in Figure 2-19. Examination of the plots for each sequence reveals key features of rate fluctuation and provides an estimate of significance for each feature. For example, in Figure 8 a significant (p < 0.1) burst of seismicity can be seen lasting 0.3 years and ending 2.4 years before the (M 5.4) mainshock. In Figure 6, a 2-year aseismic interval ending 3 years before the 1980 Livermore, Ca., (M 5.9) mainshock has a value of beta of -2.6. Despite it apparent prominance, this interval is not statistically significant (Table 5) owing to the small number of events in the sequence. In Figure 4, a low rate interval at Coyote Lake, Ca., occurring between 1977.3 and 1979.0 is also seen to be statistically not significant. In order to account for the fact that this interval begins at the same time as a magnitude correction applied to the catalog to correct for a change in network operation, Wyss and Burford (1985) suggest that the correction may be inadequate. However, in view of the fact that this low-rate interval represents an expected rate fluctuation (Table 5) for a homogeneous Poisson process, such a contention is unwarrented. In Figure 7, seismicity at Parkfield, Ca., is seen currently to be in a state of quiescence with varying significance depending on the length of interval considered. The most significant short contemporary interval of quiescence at Parkfield is the 14-month interval 1984.9-1986.1 (p < 0.05). Extreme values of beta for longer

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intervals, such as 1979-1986 (beta = -5.7), are not interpreted because these intervals excede in length the background to which they are compared. Instead, the high-rate interval in 1975-1976 is interpreted as a significant anomaly (beta = 6.7), although the distinction is admitedly philosophical.

For all the pre-mainshock sequences, intervals with anomalous rate
(p < 0.10) were identified by inspection of the calculated values of beta plotted in these figures. The pattern of anomalous intervals defined for each sequence by sojourns of the beta statistic at the 0.10 level are summarized in Figure 20. Apparently some sequences are always non-
extreme in rate, while others spend little or no time near their mean rate and alternate from one extreme to the other. A small subset of pre-mainshock sequences display precursory quiescence (sequences 6, 10, 16); others display high-rate precursory seismicity (sequences 15, 17, 21, 23). Most sequences exhibit normal-rate seismicity in the intermediate term before the mainshock. The longest and most numerous of the pre-mainshock sequences, that for the Morgan Hill 1984 earthquake, is uniform in rate for at least 10 years before the mainshock. The main result of ths section, then, is that in general, no obvious, simple pattern in intensity fluctuation is apparent by visual inspection of the collection of sequences in Figure 20.

2. Complete seismic cycles at Bear Valley, Ca.

Application of the beta statistic to the sequences of seismicity in the source regions of five sets of repeat events reveals that the sequences are generally consistent with a homogeneous Poisson process (Table 3). Only one interval - a short burst of seismicity occuring in sequence A in late 1976 - violated the beta statistic at the 90 percent confidence level. Application of the Kramer-von Mises, Kolmogorov-Smirnov, Anderson-Darling and maximum-beta tests also show these sequences to be generally uniform in rate. In fact, this group of sequences is the most uniform of the groups considered in this study.

Although each sequence, measured independently, failed, under some or all
of the tests described above, to reject the null hypothesis of a uniform process, it is still possible, with the method of kernel density estimation, to explore the sequences qualitatively for evidence of a characteristic or common intensity pattern. Average intensity estimates for optimal kernel length determined by the method of least squares cross-validation (Matthews and Reasenberg, this volume) are shown in Figure 21. The bandwidth chosen by this method minimizes a data-based estimate of the mean squared error between the true and estimated intensities. The form of the variations in estimated intensity for the five source volumes varies from sequence to sequence, with no strong pattern apparent. Intensity in the sequences for the four southeasternmost sets (A to D) is below average in the first year or so. These earthquake deficits, however, may be an artifact arising from the identification of the aftershock cluster associated with the 1972 earthquake sequence (Ellsworth, 1975). The cluster includes 1800 earthquakes and spans the interval 720222-730620 within the approximate latitude range 36°25' to 36°40'. Owing to the fact that the algorithm used to identify clusters cannot distinguish between aftershocks and normal seismicity, removal of such a massive cluster could account for the apparent earthquake deficits in the early part of sequences A through D. Stacking the average intensity estimates over a time base normalized to the length of the seismic cycle reveals the

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presence of a small sytematic increase in intensity immediately prior to the second event of the pair (Figure 22). The stack of cross-validated estimated intensities reveals the duration of the pre-mainshock increase to be not more than 6 percent of a seismic cycle, or approximately 9 months or less. An apparent decrease in seismicity rate during the first third of the composite seismic cycle may not be entirely caused by over-removal of aftershocks of the 1972 aftershock sequence, as discussed above. While this feature is farily subtle and only shows up in the composite intensity curve for five sequences, it continues beyond the end of the cluster and may indicate a real post-mainshock diminution of seismic activity in the source zones reflecting a state of lowered stress there in the beginning of the seismic cycle. In summary, the seismicity in the source volumes of the five repeat earthquakes is essentially uniform. However, a relatively low rate, common among the five sequences, in the first third of the cycle, may be associated with a reduction of stress in the source zones after the first mainshock. A small increase in rate immediately prior to the end of the cycle may be a subtle manifestation of the foreshock process.

3. Measurements of nonuniformity in intensity.

Comparisons with respect to four statistics sensitive to the nonuniformity of observed intensity revealed systematic variations in nonuniformity among the groups of sequences considered (Table 3). When considered collectively, the pre-mainshock sequences are more nonuniform in rate than the seismicity along the creeping section of the San Andreas fault, from Bear Valley to Stone Canyon. There is no apparent difference in the degree of nonuniformity of seismicity prior to mainshocks in central California and central Japan. Nonuniformity in seismicity rate was lowest among the sequences selected from the source volumes of repeat earthquakes in Bear Valley.

4. Creeping section of the San Andreas fault.

A high degree of agreement exists among the various statistical measures
of rate nonuniformity, described above, for each of the ten sequences studied in this section (Table 3). Generally, sequences that rejected the null model of a homogeneous Poisson process (volumes 382, 386, 403 and 406) did so for all four statistical tests, while most of the other sequences are consistent with the null model in all four tests. On average, the sequences in this group are more uniform than the pre-mainshock sequences, and less uniform than the source volume sequences in Bear Valley.

The five intervals of quiescence listed in Table 1 of Wyss an<i Burford
1985) are clearly detected by beta (Figures 23-27). Based on their own estimates of significance of contemporary quiescence in volumes 386 and 382, Wyss and Burford (1985) predicted the occurrence of two future earthquakes. While the high values of significance attached to their observation result from a mathematical error (Matthews and Reasenberg, in press), the present study also identifies these or similar intervals as being deficient of earthquakes, albeit at lower confidence levels (Table 4). We found the intervals of quiescence in volumes 386 and 382 to be the two most significant in this group, with the probability that each would occur by chance in a homogeneous Poisson process less than .03 and .29, respectively (Table 4). The successful prediction in Wyss and Burford (1985) was for volume 386, the only volume for which we find significant quiescence.

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DISCUSSION AND CONCLUSIONS

Several limitations in the scope of this study should be recognized before the results can be reasonably interpreted. Some limitations are inherent in the problem at hand and ultimately limit the utility of any investigation of seismicity rate fluctuations. Others are practical for this particular study and could be mitigated or eliminated in subsequent investigations. 1) By considering the multidimensional distribution of seismicity in one dimension (time), we have given up some of the available information. While acknowledging that a multidimensional study is potentially a more powerful tool for assessing seismicity patterns, we choose here to consider only fluctuations in time for two reasons. First, we wish to model our study after other recent studies of seismic quiescence so that our results might be relevent as a baseline. Second, while most earthquake catalogs are sufficiently accurate in origin time, many include relatively large uncertainties in hypocentral determinations and magnitude estimates. By simplifying the treatment of earthquake position and, to some degree, magnitude, results from widely differing catalogs can be directly compared.
2) Only 37 earthquakes from 3 catalogs were studied. A large number of events, over a wider magnitude range, drawn from world-wide and other regional and local catalogs should supplement this study. 3) In all cases except the 5 sets of repeat events in Bear Valley, only a small fraction of the seismic cycle is represented by the catalog data. In addition, rate fluctuation patterns of duration greater than, say, one-half of the catalog length, cannot be detected by the present study with confidence because the background seismicity rate is estimated from the data. Hence, except for the Bear Valley cases, rate fluctuations lasting only a few percent of the complete seismic cycle are considered. 4) Man-made rate changes in the catalog introduced by a variety of mechanisms proposed by Habermann (1986) may be contributing to the observed variance in seismicity rate. (It is considered unlikely that any real rate fluctuations have been cancelled out by coincident and opposite man-made changes. Hence, the effect of the man-made changes is probably to increase the apparent variance in rate.) While the presence of some systematic, position- and time-dependent errors in estimated magnitude for earthquakes detected by Calnet have been established by Habermann (1986) and many others may exist as well, a complete set of prescribed magnitude corrections is not yet available. Application of such a set to the California data, when it becomes available, would result in improved resoltuion in detection of rate fluctuation. 5) Second- and higher-order moments in the seismicity - i.e., relationships involving two or more events - may not be detected by this study. For example, if the rate of clustering were to change systematically in advance of mainshocks, this study, having removed a substantial fraction of the clustering, would probably fail to detect the change.

Certain measures were taken in this investigation to assure uniform treatment of the data. Seismicity prior to mainshocks was selected with only a simple (circular or spherical) region centered on the mainshock; use of complexly shaped regions based on specific knowledge of local seismotectonics was avoided. This approach has been criticized on the grounds that it ignores possibly relavent information such as the location of the fault rupture associated with the mainshock. However, because such specific knowledge is unavailable before the mainshock, a method for earthquake prediction may not make use of it. On the other hand, information derived from the catalog up to

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the time of the mainshock, such as the location of faults and fault complications infered from the spatial distribution of seismicity, may properly be used (Duncan Agnew, personal communication, 1986). Some critics have argued that our circular and spherical regions may include off-fault seismicity that "dilutes" the quiescence effect. We reply that we do not know where to expect quiescence and have no reason to favor either on-fault or off-fault seismicity. Indeed, plausible mechanisms for seismic quiescence include both whole rock (volumetric) and fault zone (areal) related dilatancy hardening (Scholz, this volume). Furthermore, some convincing cases of precursory seismic quiescence include observations not restricted to the rupture zone, extending over a considerable region (Ohtake, 1980; Wyss, 1986). In the absence of either restricting models or definitive observations we have chosen the simplest definition for test volumes. Nevertheless, specification of the spatial extent of the hypothetical quiescence is a free parameter of the present analysis, and other choices may alter the result.

Seismicity was selected from each region only for the magnitude range
above the completeness level. (An exception is the JMA data which were selected below the completeness level.) Aftershocks were removed from the CALNET and NRCDP catalogs by the same algorithm. However, the algorithm failed to remove all aftershocks from the JMA data, and some aftershocks were presumably included. The JMA catalog results are, therefore, possibly unreliable and should not be interpreted with full weight.

While these measures may make interpretation less difficult, the question still remains as to how generally the results should be interpreted. For magnitude 5.5-6.5 mainshocks we have found a broad and reliable body of evidence in the CALNET and NRCDP catalogs, and less reliable evidence in the JMA catalog, that precursory quiescence in the intermediate term is not a generally observed pattern of seismicity rate fluctuation. That is, intervals of low rate often occur, but not in a simple, systematic pattern prior to mainshocks. The hypothesis that precursory quiescence systematically or preferentially occurs prior to mainshocks in this magnitude range is not supported by the present results. The few cases displaying precursory quiescence may be adequately explained by random occurrence.

In view of the foregoing discussion, what should be said about the body of published studies that show, in some cases, rather convincing evidence for precursory seismic quiescence To what extent does the negative result of the present study conflict with these First, for mainshocks from other regions, or with magnitudes outside the range of magnitudes studied here, this study does not directly conflict. Furthermore, within the present magnitude range, the results of the present study are not inconsistent with a finding of a strong pattern in rate fluctuations in any particular case. Indeed, we have demonstrated the existence of many such patterns here, and have shown the occurrence of anomalous-rate intervals to be substantially greater than that expected for a homogeneous Poisson process. While this study finds no simple, prefered pattern of precursory quiescence, it demonstrates that rate fluctuations are a common precursory feature of seismicity in the source region.

The present negative result does not necessarily conflict with the fact
that numerous cases of precursory seismic quiescence have been reported in the literature. The following model, illustrated in Figure 28, generally applies

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to any proposed earthquake precursor. Let the set of instances of a particular proposed precursor pattern (or anomaly) be represented by A. Let the set of mainshock earthquakes be represented by E. The intersection of these sets, S, is the set of successes of the pattern to predict an earthquake. Suppose the complete set of anomalies is known. Let the number of anomalies be n(A), the number of earthquakes be n(E), and the number of successful predictions be n(S). We define two parameters:

n(S)
V =
nTO

n(S)
and R = ___
HnTFT

V is the "validity" of the proposed pattern. When V is large, it is valid to predict an earthquake based on the occurrence of the anomaly because a large proportion of anomalies are followed by an earthquake. R is the "reliability" of the proposed pattern. When R is large, earthquakes are reliably predicted by anomalies because a large proportion of earthquakes are preceded by an occurrence of the pattern. Clearly a desirable property of a proposed predictor is that both V and R be large. If V is small, a large proportion of the predictions are false alarms. If R is small, a large proportion of the earthquakes go unpredicted. The present investigation has approached the question of precursory quiescence from the side of the earthquakes. That is, we started with a set of earthquakes and found R to to be small. On the other hand, the body of reported cases of successful predictions by the quiescence hypothesis is represented by S. Although the size of A is not known, let's suppose V is not very small, say as large as 0.5. (i.e., the quiescence hypothesis is valid). The present study does not conflict with this supposition. It concludes that R « 1, indicating that the quiescence precursor is not reliable; most mainshocks are not systematically preceded by it. A corresponding result for foreshocks was obtained by Von Seggern, et al. (1981), who found for 510 mainshocks (M > 5.8), that foreshocks were the exception rather than the rule.

In terms of the foregoing discussion, the validity of seismic quiescence
as an earthquake precursor has not been tested in the present investigation. Even if most earthquakes are generally not preceded by quiescence, establishing that mainshocks follow a substantial proportion of all occurrences of quiescence would be an important and immediately useful step in earthquake prediction. To do this would require a systematic search over a suitable, complete catalog with a single definition and methodology for detecting quiescence. (If in the process, some of the parameters used to define quiescence or methods to detect it are determined from the data, then the usual precautions of withholding a portion of the data for purposes of cross-validation would be indicated.) Identification of a complete set of occurrences of seismic quiescence would be a valuable contribution toward establishing the validity of the hypothesis.

In summary, we conclude from these results that:

1. The nonuniformity in seismicity rate in the regions immediately

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surrounding future mainshocks is greater than expected for a homogeneous Poisson process. Rate fluctuations that depart significantly from a uniform random model are common features of pre-mainshock seismicity. Pre-mainshock seismicity rate is highly nonuniform for some sequences and uniform for others. Seismicity rate generally varies less within selected areas along the creeping section of the San Andreas fault than in the pre-mainshock source regions studied.

2. A weak signal in the variation of rate of seismicity was found in the source regions of five repeat earthquakes in the vicinity of Bear Valley, Ca., over their entire seismic cycles. The signal, which could be seen only by combining the results for the five sequences, consists of a) a small increase in rate in the 6 to 9 month period at the end of the cycles possibly related to the foreshock process (before the second of the repeat earthquakes), and b) a period of low rate in the first third of the seismic cycle. The latter observation may reflect a state of stress relaxation in the source volume immediately after the mainshock.

3. A simple, widespread (spatially stationary) pattern of precursory seismic quiescence is not apparent, in the intermediate term, in the set of cases considered. The body of reported cases of precursory seismic quiescence may satisfactorily be explained by a model in which random rate fluctuations are preferentially reported when they conform to the hypothesis (i.e., successes are reported preferentially over failures). While precursory seismic quiescence clearly occurs and may be an important observation for understanding the physics of the earthquake generation process, no evidence was found that the phenomenon satisfies basic criteria qualifying it as a reliable predictor of future earthquakes.

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TABLE 4

Quiescent Intervals Identified by Wyss and Burford (1985) Quiescent Intervals Identified in This Study

VOLUME
I.D.

361
372
382
386
401

State of
Interval

Dec 1980
Jun 1982
Feb 1983
Jun 1983
Jul 1978

Duration End of
(Years) Interval
(1)
0.9
2.5
1.9
1.5
0.8

Oct
Dec
Dec
Dec
May

1981
1984
1984
1984
1979

Estimated Significance
of Observation

p < 0.01
p < 0.01
p < 0.01
p < 0.01
p < 0.01

Start of
Interval

Feb
Jun
Feb
Dec
Apr

1979
1983
1983
1983
1978

Duration End of Estimated Signi
(Years) Interval of Observat

3.0
1.5
2.7
2.0
0.9

Feb 1982
Jul 1984
' Oct 1985
Dec 1985
Mar 1979

P
P
P
P
P

< 0.50
< 0.40
< 0.29
< 0.03
< 0.43

Notes

TT]The Wyss and Burford study included data up to December, 1984.

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CO

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FIGURE CAPTIONS

Figure 1. Locations of the five sets of repeat earthquakes on the San Andreas fault near Bear Valley. Cirices of 2-km and 3-km radius shown surrounding the centroid of hypocenters in each pair (top) indicate the spherical source volumes used to define the seismicity sequences studied. Only results for the 3-km radius spherical volumes are presented.

Figure 2. Seismicity for the 10.5-year pre-mainshock interval prior to the April 24, 1984 (M 6.2) earthquake at Morgan Hill, Ca. is shown in a variety of ways, a) Contoured values of the beta statistic are shown for all possible intervals on a grid of interval end-times and interval durations with increments of 30 days. Each value of beta is shown at a position corresponding to the interval ending time and duration. Values of beta are calculated only for intervals that begin and end entirely within the sequence (represented by the lower right triangular portion of figure). Contour level spacing of 3.8 equals the one-sided 0.1 level critical point for beta. Thus, intervals with scores of beta less than -3.8 are quiescent at the 90 percent confidence level. For no interval in this sequence does beta excede 3.8 in absolute value. Thus, the sequence fails to reject a uniform Poisson process at the 90 percent confidence level. Local relative extrema of beta indicate intervals with relatively extreme rates. For example, the circled value of 2.6 corresponds to an interval with duration 1.9 years and ending at 1979.5. This interval is not particularly extreme in rate; the value of beta of 2.6, while the highest for this sequence, is not significant at the 50 percent level (Table 5). Seismicity for the 10 years before the mainshock at Morgan Hill was apparently highly uniform, b) Seismicity for the same sequence shown as a "stick plot". Each line represents an earthquake with origin time indicated by the time scale in (a) above. Height of line corresponds to event magnitude. The interval represented by the circled relative maximum in beta of 2.6 in (a) is indicated, c) Cumulative number plot for the sequence, d) Estimates of intensity (rate) of the sequence calculated by the method of kernal density estimation. Smoother curved line is estimated intensity obtained with the optimal averaging kernel. Less smooth curved line is estimated intensity obtained with kernel length 0.5 times optimal. Lengths of the averaging kernels are indicated in the lower left corner. Straight broken line represents uniform Poisson intensity. The interval represented by the circled relative maximum in beta of 2.6 in (a) is indicated. A gradual increase in intensity over the length of the sequence is observed in the Morgan Hill sequence.

Figure 3. Seismicity for the 12-year pre-mainshock interval prior to the 1986 (M 5.9) earthquake at North Palm Springs, Ca. See Figure 2 caption for further explanation.

Figure 4. Seismicity for the 5.6-year pre-mainshock interval prior to the 1979 (M 5.9) earthquake at Coyote Lake, Ca. See Figure 2 caption for further explanation.

Figure. 5 Seismicity for the 9.5-year pre-mainshock interval prior to the 1983 (M 6.7) earthquake at Coalinga, Ca. Significant rate change beginning in 1980 is associated with a swarm that partially extends into the study area (Eaton, 1985). See Figure 2 caption for further explanation.

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Figure 6. Seismicity for the 6-year pre-mainshock interval prior to the 1980 (M 5.9) earthquake at Livermore, Ca. The small number of events in the sequence results in the failure of beta to achieve a significant level for the 2.1-year aseismic interval (1975 to 1977). This example illustrates how classical visual inspection of seismicity in (b) or (c) can misperceive an insignificant feature as a very significant one. See Figure 2 caption for further explanation.

Figure 7. Seismicity for the contemporary 12-year interval of seismicity surrounding the 1966 epicenter at Parkfield, Ca. This region is currently in a state of significant seismic quiescence. See Figure 2 caption for further explanation.

Figure 8-13. Seismicity for the pre-mainshock intervals prior to six (5.3 < M < 6.4) earthquakes in the Tokai-Kanto district of central Japan, obtained from the NRCDP catalog. See Figure 2 caption for further explanation.

Figures 14-19. Seismicity for the pre-mainshock intervals prior to six (6.0 < M < 6.7) earthquakes in central Japan, obtained from the JMA catalog. See Figure 2 caption for further explanation.

Figure 20. Summary of anomalous rate intervals identified with the beta statistic for the pre-mainshock sequences. Each sequence is represented by a horizontal bar on a common time scale aligned at the right margin with the time of occurence of the mainshock. Intervals of anomalously high (low) rate, defined as sojourns of beta at the 90 percent confidence level, are represented by vertical (diagonal) hatchure. Intervals with normal rate are represented by solid bar. Representation of the rate fluctuations in each sequence by this simple display of sojourns of beta facilitates a search for a common pattern of rate fluctuation prior to the mainshocks. Numbers on right identify sequences by reference to Table 6.

Figure 21. Seismicity in the source region of five repeat events on the San Andreas fault near Bear Valley, Ca. during the intervals between repeat events. Stick-plots indicating time sequences of seismicity (top) are annotated above with the magnitudes of the main (repeat) events defining the sequence. Corresponding cumulative number plots are shown directly below. Averaged intensity of the processes underlying the observed seismicity, estimated with cross-validated least-squares averaging kernels (see text) are shown in bottom portions of figure (curved lines) together with the rate corresponding to a uniform Poisson process for the sequence (broken lines). Averaging kernel lengths are indicated in lower left corners. All five sequences are uniform in rate. Sets A, B, C and D show below average rate during the first year of the sequences, possibly owing to the removal of the aftershock cluster for the 1972 earthquake sequence.

Figure 22. a) Composite (stack) of the five average intensity estimates shown in Figure 21 on a common timebase normalized to the length of the seismic cycles. Vertical scale is normalized rate. Lowest rate in composite occurs in the first third of cycle, and may reflect a real seismicity decrease associated with stress relaxation following the first mainshock. Highest rate in composite occurs at the end of the cycle, and may reflect a weak foreshock process, b) Composites of estimated intensity functions obtained with averaging kernels of length 6 months, 12 months and 24 months.

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Figureg 23-27. Seismicity in selected volumes along the creeping section of the San Andreas fault. Significant quiescence anomalies were reported for each volume by Wyss and Burford (1985). a) Figures from Wyss and Burford (1985) indicating departure of indicator function AS(t) used to identify the anomaly, b) Cumulative number plot obtained in this study for M > 1.7. Differences between these and the corresponding cumulative number plots in (a) are due to differences in the aftershock removal procedures employed in the two studies. The intervals corresponding to the ones identified as anomalous by Wyss and Burford are indicated along with the significance levels obtained in the present study (see Table 4). For volume 386 (Figure 23) good agreement exists in the identification of a significant interval of quiescence beginning about 1984. Data for the Wyss and Burford study ended in December, 1984, while the present study extends to December, 1985. Nevertheless, generally good agreement is found for the onset time and duration of these quiescent intervals. However, the present study finds all the intervals except the one in volume 386 to be not statistically significant. This discrepancy arises from an overestimation of the significance of fluctuations identified by AS(t) in the earlier work (see text).

Figure 28. Venn diagram illustrating the relationship between the set, (A), of occurrences of a pattern (or anomaly) proposed for earthquake prediction, and the set, (E), of earthquakes elegible for prediction by the anomaly. The intersection, (S), of these sets implies the acceptance of some correlative method to relate anomalies and earthquakes. Then, the size of (S) relative to the size of (A) and (E) defines, respectively, two independent measures (V and R) of the predictive power of the proposed pattern or anomaly. See text for further explanation.

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COMPOSITE OF 5 COMPLETE SEISMIC CYCLES

3-KM RADIUS SOURCE REGION

OPTIMAL BANDWIDTH

0

24 MO\

12 MO-

6 MO

NORMALIZED TIME

0
NORMALIZED TIME

FIGURE 22

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Anomalous Seismicity in the San Diego Coastal Region

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Abstract
This short note documents recent seismic activity in the coastal region adjacent to San Diego. Seismic activity in the region has increased dramatically since 1983 when compared with previous historic activity. Similar increases in seismicity, that occurred prior to significant earthquakes in California, are also documented. It is speculated that the tectonic style in the continental borderland off of San Diego is very similar to that in the Basin and Range province.

Introduction
Despite the fact that large linear bathymetric features in
the continental margin of southernmost California have long been suspected to be related to major active faults (Clarke et al., 1985), earthquake activity in the vicinity of San Diego has historically been very low. However this situation has changed rather dramatically in the last several years. A magnitude 5.3 earthquake, commonly referred to as the Oceanside earthquake, occurred at 13:46 GMT, 13 July 1986 about 70 km northwest of San Diego along the northern end of a major northwest-trending escarpment known as the Thirty-Mile Bank (see Figure 1). This earthquake was preceded by over a year of increased activity in the epicentral region and the aftershock sequence has also been quite strong. In addition to the Oceanside sequence, San Diego has also experienced a sequence of five earthquakes in the magnitude 4 range that have occurred both in San Diego Bay and slightly offshore within the last year and a half. For long-time residents of San Diego, this earthquake acitvity has certainly seemed unusual. What is the significance of these sequences; are they likely to continue; and what is the prospect for even larger earthquakes within the next several years?

Recent Seismicity
Recent seismicity in the San Diego region is shown for the periods 1978 through 1984 and 1985 to Sept. 1986 in Figures 1 and 2, respectively. A time-distance plot of seismicity as projected onto the line C-D is shown in Figure 3 for the period from 1978 through Sept. 1986. A number of clusters of activity can be distinguished; one large one to the north (Oceanside earthquake), and two (maybe more) clusters in the San Diego region. Clearly, the rate of seismicity since 1983 is far higher than that from 1978 to 1983. There are several indications that these clusters may be temporally related on several time scales. The most obvious correlation is the appearance of these clusters within a two-year period. A second temporal coincidence can be seen on a time scale of days. For instance, the Oceanside earthquake in July 1986 was preceded by three hours by a magnitude 3 earthquake

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in San Diego Bay. Within the next week, clusters of seismicity were active about 10 km west of San Diego. A similar coincidence in June 1985 prompted an official notification from the USGS to the State of California of a potential geologic hazard. In that case, a M 3.9 earthquake in the Oceanside region was followed within two days by a sequence of earthquakes in San Diego Bay having magnitudes of 3.9, 4.0, and 3.8. Although it may be that these coincidences are merely fortuitous, they are certainly disconcerting when experienced in real time.
The recent activity in the San Diego coastal region can be viewed from a broader perspective in Figure 4. All seismicity located using the Caltech-USGS seismic network since 1978 is shown. The recent clusters of activity appear to lie along and to define a rather broad seismicity lineation running from the San Miguel fault in northern Baja to the Santa Monica Bay. A second broad seismicity lineation can be seen to the west running from Ensenada to Santa Barbara. At this point, I am unaware of any mapped throughgoing fault systems that parallel these apparent seismicity lineations and it may be that the lineations are caused by a fortuitous alignment of earthquake clusters. Focal mechanisms for the 13 July 1986 M 5.3 earthquake and for the M 3.8 earthquake that occurred in the San Diego sequence of June 1985 are shown in Figure 5. Focal mechanisms for the San Diego earthquake appear to be consistent with right-lateral faulting on northerly-trending faults. These mechanisms would be consistent with the orientation of the Rose Canyon fault system. The focal mechanism of the Oceanside earthquake is somewhat problematic; the focal mechanism from local readings from the southern California network yields oblique-right-lateral slip on a north-trending plane (Egill Hauksson and Lucy Jones, personal communication). However, long-period teleseismic P-waves seem to be more consistent with a N46W-striking, 52E-dipping, thrust fault (John Nabelek, personal communication). Furthermore, many of the aftershocks for which mechanisms have been determined using local array data also have significant thrust components (Lucy Jones, personal communication). Thirty-Mile Bank, the major bathymetric feature of the area, trends to the northwest. At this time, the orientation of the Oceanside earthquake is unresolved. Time-distance plots of seismicity projected onto the line A-B
of Figure 4 and for the periods 1932 to present (magnitude > 3.8) and 1978 to present (all magnitudes) are shown in Figures 6 and 7, respectively. A noticeable paucity of significant seismicity in the San Diego region can be seen for a period from the mid 50's through the mid 10's. The onset of activity seen in Figure 6 is due to events to the southeast of San Clemente island. The persistent clustering nature of the recent seismicity can be seen in Figure 7.

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Comparison to Other Regions
One of the most disconcerting features of the recent San Diego seismicity is its persistence and its increasing rate with time. When experienced in real time, it is far easier to get excited by increasing seismicity than by seismic quiescences. However, there are several clear examples in which persistent, increasing seismicity over the time scale of months has been accompanied by potentially damaging earthquakes. A number of rectangular subregions of California are defined in Figure 8. In Figure 9, the cumulative number of earthquakes larger than magnitude 3.5 is given as a function of time since 1932 for each region. This same function is also plotted for the entire Caltech catalog. Although a careful study of the procedures used to construct this catalog would be necessary to assess the completeness of the catalog, I believe that the catalog is reasonably complete at the magnitude 3.5 level (Hileman, 1978). Clear changes in the seismicity rate can be seen in 1952 and 1979 for the entire catalog. I am not aware of any procedural changes that would have produced the lower overall seismicity rate for the period from 1952 to 1979, and one speculation is that this change is physically related to the occurrence of the M 7.7 1952 Kern County earthquake (Raleigh et al., 1982). The overall catalog seismicity rate increased again in the period 1979 to present principally because of earthquake sequences in Imperial Valley, Long Valley, Coalinga, and most recently, North Palm Springs and Oceanside.
Inspection of the cumulative seismicity plots for selected subregions (Figure 9) reveals that periods of increasing seismicity have preceded several significant earthquakes. Very clear seismicity increases can be seen in both the Mammoth and Coso regions (Boxes A and C, respectively). Three magnitude 6 earthquakes occurred in 1980 in the Mammoth region following a period of increasing activity that began in 1978 (Ryall and Ryall, 1981) . The recent increase in activity in the Mammoth box is distributed over a very large region and high levels of seismicity have persisted in this region at least until the present. Seismicity in the Coso box also experienced a steady increase during the period from 1981 through 1983. The largest events in this region have been in the magnitude 5 range and were preceded by about a year of increased seismic activity.
Seismic activity in the Imperial Valley region (Box F) was also clearly higher for a period of about 5 years before the magnitude 6.5 Imperial Valley earthquake in 1979 (Johnson, 1979). Finally there was a noticeable increase in seismic activity in the area of the 1983 M 6.5 Coalinga earthquake (Box B) and a magnitude 5.5 earthquake occurred just to the north of and six months prior to the Coalinga mainshock (Eaton, 1985). The seismicity in the San Diego region (Box G) shows a very clear seismicity increase that began in about 1983. This increase is very similar to the increased seismicity that was experienced in the Mammoth and Coso regions.
Although I have shown several examples in which seismicity increased over the time scale of months to years before several significant earthquakes, there are also many examples of larger

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earthquakes for which seismicity increases did not occur. There is no evidence that seismicity increased before the M 6.5 San Fernando earthquake in 1971 (Box D) or before the M 6.5 Borrego mountain earthquake on the San Jacinto fault in 1968 (Box E).

Discussion
With the exception of the Coalinga earthquake, the previous examples of increased seismicity occurred in regions of strike-slip faulting that are accompanied by crustal spreading. A natural question arises; is the recent seismicity in the San Diego region also related to strike-slip faulting in a region of crustal spreading? Inspection of the bathymetry in the continental margin south of the Los Angeles Basin and north of Ensenada, Mexico reveals a very broad region (more than 200 km wide) of northwest-trending basins and ranges. In fact, the physiography of this region is very similar to that of the Owens Valley, Panamint Valley, and Death Valley. The vertical relief seen within this continental margin is as much as 2,000 meters and the physiography of this margin is unique when compared with any other continental margin. Humphreys and Weldon (1986) have suggested that approximately 5 to 10 mm/yr of northwest-trending, right-lateral shear deformation is occurring in the continental borderland. This shear deformation may also be accompanied by an undetermined amount of crustal spreading (Gene Humphreys, personal communication).
There are indications that the Basin and Range province of eastern California and Nevada have a distinctive style of seismicity in which regions undergo long periods of seismic quiescences (hundreds or thousands of years) that are interrupted by periods of high regional seismicity (Wallace, 1985). If this is also the case for the continental margin off of San Diego, then we may be seeing the beginning of a prolonged period of higher seismicity. Although I have not been able to quantitatively give any prospects for future activity, I believe that there is an even chance that we have not seen the last of the seismic activity in this region.

Acknowledgments
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San Diego Section

VENTS URN 85 TO RUG 86)

LESS THRN 3.0

3.0 TO 3.99

U.O TO 4.99

5.0 TO 5.99

LflRGER THflN 6.0

Figure 1. All events in the Caltech catalog from Jan. 1985 through Aug. 1986. Many events after the 13 July 1986 earthquake are not yet included in the catalog. Seismicity northeast of the Elsinore fault is not plotted. Time-distance plot for seismicity in the box and projected onto the line C-D is found in Figure 3.

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LESS THflN 3.0

3.0 TO 3.99

4.0 TO 4.99

5.0 TO 5.99

LflRGER THflN G.O

Figure 2. Same as Figure 1, except for the time period from Jan 1978 through Dec. 1984.

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South Coast Section

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i-'^V . . -^ . / i. . . - K-;-,-^

LESS THflN 3.0

3.0 DJL3.99

4.0 TO 4.99

5.0 TO 5.99

6.0 flND LflRGER

120

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Figure 4. All events in the Caltech catalog from Jan. 1985 through June 1986 (about 100,000 events). Time distance plots for seismicity in the box are given in Figures 6 and 7.

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28 miles SM of Oceanslde M=5.3

13 July 1986

San Diego

Figure 5. Focal mechanisms of the 13 July 1986 M 5.3 Oceanside earthquake and the 17 June 1985 M 3.8 San Diego earthquake from first-motion data from the Caltech-USGS telemetered array (Lucy Jones and Egill Hauksson, personal communication). Long-period teleseismic P-waves indicate a thrust mechanism for the Oceanside earthquake (dotted line; John Nabelek, personal communication).

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EQ_CRT RLL 9
38

37

21

LESS THRN 3.0
3.0 UJL3.99
4.0 TO 4.99

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120

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119 118 117 116 115

Figure 8. Definition of boxes used in Figure 9 together with seismicity in the Caltech catalog from Jan. 1978 to June 1976.

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1932

c 0) > 0) "o
6

Caltech Catalogue
CF r/g--i:s ,">;.;:______

Mammoth

Coalinga 1988 1932

B ...

CosQ 1988

C

Figure 9. Cumulative number of earthquakes larger than magnitude 3.5 plotted versus time since 1932. The entire Caltech catalog is plotted in the upper left and the other subregions are defined in Figure 8.

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PLAUSIBILITY CRITERIA FOR INTERMEDIATE TERM EARTHQUAKE PREDICTION

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ABSTRACT

Geophysical anomalies can be scientifically useful and justify public action only if previous examples satisfy the following criteria:

1. Anomalies must be observed simultaneously at several sites.
2. Coseismic anomalies must be observed at the same
sites as the preseismic anomalies.
3. The magnitude of the anomaly must bear a reasonable relationship to epicentral distance.

Few of the stress, strain, tilt, electrical, magnetic, and geochemical precursors reported in Japan, China, or the US satisfy any of these criteria, and none satisfies all three. Explanations include the following: (1) observing stations are too sparse, (2) the stress fields from preseismic events have nodes at the surface, and (3) the extreme sensitivity of some sites overcomes the effect of epicentral distance. Each of these explanations involves a hypothesis, which must be tested before we can move from speculation to science. The first two hypotheses may be tested by greater station density. However, we do not presently have affordable means to measure most geophysical phenomena accurately enough, so basic research into measurement techniques is needed before we can productively study earthquake precursors. The third hypothesis should be testable using existing data in many cases; if a particular site is exceptionally sensitive to precursory strains or stresses, it should be similarly sensitive to tidal or other predictable effects.

INTRODUCTION

For many years, earthquake prediction research has been in an exploratory phase. We have searched for possible relationships between coming earthquakes and many geophysical quantities, hoping that anomalous variations in these observable would somehow indicate when the earth is near its breaking stress. We now have a long catalog of phenomena which in some circumstances show anomalous behavior before some earthquakes. But in every case the relationship between the geophysical anomaly and the ensuing earthquake is inadequately understood or even dubious. Before we can ask public officials to act on our predictions, we must move beyond the stage of cataloging

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curious anomalies. We need to achieve a scientific understanding of the earthquake process and its relationship to geophysical quantities. Application of the three criteria listed in the abstract could help us focus our attention on phenomena which can really be related to tectonic stress, and to design experiments adequate for scientific progress.

SIMULTANEOUS OBSERVATIONS

The first criterion, that of simultaneous observation, is a test that the observed anomaly is not purely an instrumental aberration, and that the anomaly is somehow related to stress. Simultaneous observations will also be necessary if we are to understand the cause of the anomaly. We would not think of modeling static gravity or magnetic anomalies based on one measurement, nor can we infer the cause of an earthquake precursor from one measurement. To obtain simultaneous observations of the same precursor at several sites, we must plan in advance to have closely spaced observation sites. If a network of sites at 10 km spacing records some possible precursors at just one site on each occasion, then a smaller spacing is required. Obviously for a given budget, increasing station density implies accepting a smaller geographical coverage. It is a sacrifice that must be made.

A geomagnetic variation was observed on a network of proton-precession magnetometers in central California prior to the 1974 Thanksgiving Day Earthquake. Smith and Johnston (1976) reported an anomalous variation of about 1 nT in the difference field between two sites within 20 km of the Thanksgiving day event. The anomaly started about 7 weeks prior to the event and lasted for a few weeks. After heavy data processing, Davis et al. (1980) were able to find some rather subtle variations at other stations within 44 km of the epicenter, at the same time as the first reported anomaly. As there was no coseismic magnetic variation, it is difficult to be sure that the observed anomalies were of tectonic origin, but at least they were not caused by instrumental malfunction.

Another example of apparently simultaneous precursors occurred for the 1978 West Off Izu-Oshima earthquake. Oki and Hiraga (1987, this volume) reported that many water wells within 100 km of the event reported changes in water level of up to 100 mm about 1 month before the event. Also, borehole dilatometers at Ajiro (35 km north of the epicenter) and at Irozaki(40 km southwest of the epicenter) showed variations from their long term trends beginning about one month before the earthquake. But the various anomalies actually have different starting times and waveforms. It is clear that the water level variations are not simply responding to a widespread strain variation, or else the water level records would be more similar to one

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another. Because water level variations respond in a complicated, time dependent way to strain changes (and rainfall and other phenomena) distributed over a large aquifer, some variation in response is to be expected. But in order to interpret the data and understand the cause of the anomalies, some modeling is necessary to "deconvolve" the observations at each site to recover the strain or other source of water variations. Until this is done it is too early to judge whether the water data truly represent simultaneous observations of an earthquake precursor.

COSEISMIC RESPONSE

Because earthquakes generally cause the largest strain and stress variations of tectonic origin, any measurable quantity that responds to tectonic strain or stress should respond to earthquakes. Thus a hypothetical precursor that does not show coseismic variations must be suspected of having a nontectonic origin.

Some would argue that perhaps a slow precursory stress release may actually have a larger moment than the earthquake it foretells. But geodetic estimates of moment release, which include contributions from preseismic, coseismic, and postseismic effects, generally agree with the seismic estimates to within a factor of two, and any real differences are probably caused by postseismic rather than preseismic phenomena. Johnston et al. (1986) report on extremely precise measurement of crustal strain in central California using a borehole dilatometer. They observed a coseismic strain change for a M= 3.2 earthquake, but there was no precursory strain above a level of about 0.01 nanostrain, or about 3% of the static coseismic strain change. Other observations with very accurate and sensitive strainmeters have shown similar results. Since the coseismic stress changes are larger by orders of magnitude than any credible precursory changes, a hypothetical precursor that does not respond to seismic stress changes must be viewed with extreme suspicion. Many of the quantities we measure also respond to rainfall, temperature variations, atmospheric electrical fields, and other nontectonic conditions.

It is also possible that preseismic stain changes may occur fortuitously close to an observing site, or that the preseismic anomalies might have a different mechanism which would favor preseismic observations over coseismic effects. While this could happen under special circumstances, it is unlikely to be a common occurrence. Clearly we must place our measurement sites close together enough that we can interpret both the preseismic and coseismic variations in terms of their source mechanisms if we are to make scientific progress in earthquake prediction.

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Certainly the possible geomagnetic precursor to the 1974 Thanksgiving Day earthquake would have been much more convincing if there had been a coseismic effect as well. Gladwin and Johnston (1987, this volume) report coseismic magnetic variations (without precursory anomalies) at several sites during the 1986 North Palm Springs earthquake in California. But the lack of previous coseismic anomalies shows that the relationship between stress and magnetic variations is very complicated. For this reason it will be awfully difficult to interpret geomagnetic anomalies and to make earthquake predictions from them.

Water levels varied at some wells at the time of the 1978 West Off Izu-Oshima Earthquake, but not at most of the wells reporting precursory changes. And the more precise dilatometers failed to record a coseismic signal comparable to the proposed precursory signal. The relatively large spacing between dilatometers does leave open the possibility that the recording instruments were both at nodes of the coseismic response. Only more accurate data at more closely spaced sites can resolve this problem. Further examples of possible precursors recorded on the present array of instruments may stimulate our curiosity, but they are unlikely to save lives or lead to deeper understanding.

RELATIONSHIP TO EPICENTRAL DISTANCE

Examples of proposed earthquake precursors at many tens and even hundreds of km from epicenters of later earthquakes abound in the literature(for example, King, 1986; Deng et al., 1982; Wakita et al., 1987). But in many cases stress-sensitive measurements much closer to the epicenters show no anomalous variations before or during the earthquake. How can it be that some sites far from an event respond to precursory stress changes but nearby sites don't? This question needs a convincing answer if we are to expect our governments to invest in chemical, electrical, gravity, magnetic, and other geophysical experiments to detect such distant anomalies.

One popular hypothesis is that some sites are more sensitive to stress, even by orders of magnitude, than other sites. Thus widespread but very subtle stress changes, too small to be detected at most sites, may stimulate large responses at a few sensitive sites. For example, a compliant boundary zone (that is, a fault) between two or more rigid blocks would suffer much larger strains in response to a regional stress than would sites within the rigid blocks. While there may be much truth to this reasoning, we cannot rely upon it to explain discrepant data without a rigorous test. To interpret earthquake precursor data we need to quantify the sensitivity of the site so we can infer something about the regional stress or strain field. Sensitive sites should also respond strongly to tides,

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surface waves, atmospheric pressure gradients, and other sources of long-wavelength stress variations. We need to use these calibration tools to the fullest possible extent.

Simple geophysical arguments suggest that the strain response to regional stress may vary by as much as a factor of two from site to site, but order of magnitude variations are unlikely. Seismic wave velocities are determined largely by rock compliance, and lateral variations of more than 50%, even within fault zones, are rare. Variations in tidal gravity admittance are also quite limited. Order of magnitude variations in strain sensitivity from site to site are incredible.

Of course some precursors may respond not to strain but to electrical resistivity, magnetic susceptibility, or other quantities which do have strong local variations. Use of these techniques in earthquake prediction presents substantial technical problems. First, it will be exceptionally difficult to quantify the relationship between stress changes and the measured response. Second, it may be impossible to find enough suitable sites that are both in sensitive zones, and well distributed around earthquake preparation zones.

A serious problem with sensitive sites, both for strain and indirect techniques such as geochemical and geomagnetic monitoring, is that they may also amplify nontectonic noise such as rainfall induced strains, thermal variations, etc. For most modern geophysical measurements, information is limited more by the signal/noise level rather than by sensitivity. Because much of the noise is geophysical rather than electronic, sensitive sites may not offer a significant advantage in earthquake prediction. Rather, because of their complex response to stress, they may serve to confuse rather than enlighten us.

The water level variations before the 1978 West Off Izu-Oshima earthquake offer an example of observations with a problematical relationship between response and epicentral distance. Oki and Hiraga (1987, this volume) show that some wells over 140 km from the event observed anomalous water levels before the event, while other sites within 50 km did not. The largest response came at a site 110 km from the epicenter. Before these data can be used to infer that precursory stress changes really occurred, it must be shown that the more distant sites observing large anomalies have proportionally large tidal response. Otherwise, if the largest anomalies must be discounted, the credibility of the entire dataset is jeopardized.

SUMMARY

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Earthquake prediction research is, or at least better be, at an important crossroads. We have seen enough interesting, relevant, and possibly valid earthquake precursors to know that more of the same will not be sufficient to make reliable intermediate-term earthquake predictions. We need to improve our measurement capabilities, especially for quantities such as strain and tilt that respond directly to stress, and we need to make observations at much closer spacing (possibly a few km or less) than we have in the past. We must demand that proposed earthquake precursor satisfy the three minimal criteria above. We should recognize and adopt these criteria now, so that we can design experiments that will lead to useful scientific understanding of earthquake precursors, if any exist.

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PROBABILITY OF EARTHQUAKE RECURRENCE WITH NONUNIFORM STRESS RATES AND TIME-DEPENDENT FAILURE

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ABSTRACT

Current methods for calculation of long-term probabilities for the recurrence of large earthquakes
on specific fault segments are based upon models of the faulting process that implicitly assume constant stressing rates during the interval separating earthquakes and instantaneous failure at a critical stress threshold. However, observations indicate that the process of stress recovery following an earthquake involves rate variations at all time scales in addition to stress steps caused by nearby earthquakes. Additionally, the existence of foreshocks, aftershocks and possible precursory processes suggest that there may be significant time dependence of the earthquake nucleation process. A method for determining the conditional probabilities for earthquake occurrence under conditions of irregular stressing is developed that could be useful at all time scales including those pertinent to short- and intermediate-term prediction. Used with models for earthquake occurrence at a stress threshold, the addition of variable stressing introduces a simple scaling of the conditional probabilities by stress level and stress rate. A model for the time-dependent nucleation of earthquake slip has been proposed recently that is based upon laboratory observations of fault strength. Using this time-dependent failure criterion, it is found that large but relatively short duration changes in the probability of earthquake recurrence can arise particularly following stress steps. The model suggests that variations of seismicity rates of small earthquakes in the nucleation zone of the expected earthquake directly indicate variations in probability of recurrence of the large earthquake.

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INTRODUCTION

The concept of an earthquake cycle in which elastic straining slowly restores the stress released
in a previous earthquake to a critical state preparatory for the next earthquake has long been a foundation of theories for crustal earthquakes (e.g. Reid, 1911). This concept forms an underlying basis for various methods that have been employed for estimating the time of occurrence of future large earthquakes on specific fault segments. Three methods that provide increasingly quantitative estimates of the potential for earthquake occurrence are: the classification of seismic gaps by elapsed time from the last large earthquake (McCann et al, 1979; Nishenko and McCann, 1981); calculation of the the expected time of occurrence from slip in the previous earthquake, and long term slip rates (Shimizaki and Nakata, 1980); and estimated conditional probability for recurrence of earthquake slip (Lindh, 1983; Sykes and Nishenko, 1984).
Calculation of the probability for earthquake recurrence as implemented by Lindh (1983) and Sykes and Nishenko (1984) appears to be warranted as mechanically reasonable for this critical stress model. Although the probability formulations make no provision for possible variation in stressing rates or stress steps, it is evident that stress accumulation is not necessarily a linear process. Consequently, there is the possibility of a significant time-varying component to the earthquake probability that is not treated in current approaches. Non-linear stressing could significantly alter not only the expected recurrence intervals but, also other details of the earthquake cycle that might be stress dependent (i.e. Thatcher, 1982).
Strain rate variations have been reported from tectonic regions at all time scales. At short time scales the stressing rates from solid earth tides typically exceed average tectonic stressing rates by an order of magnitude or more. Sudden stressing and stress steps are well documented to occur as result of large nearby earthquakes, intrusion events in volcanic areas and from creep events on nearby faults. Paired earthquakes, foreshock-mainshock sequences and increased earthquake activity related to intrusive events all appear to be cases where strain events increase earthquake probability.
At the longer time scales of months to years, strain events and strain rate variations, usually of unknown origin, have also been widely reported. However, at these time scales, the observations have generally been near the limit of resolution or stability of observing systems and consequently verification tends to be less conclusive. In southern California, episodes of possible large-scale non-linear straining have been reported from geodetic observations by Savage et al (1981), Savage and

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Gu (1985), Whitcomb (1980), Neill, et al. (1975) and Smith, et al. (1985). Jachens, et al. (1983) describe episodes of correlated gravity, elevation and strain changes in southern California. At a smaller scale in the vicinity of Palmdale, California, data from a two-color laser ranging array show evidence of a strain event in 1984 in which the rate of shear straining acting parallel to the San Andreas fault underwent a roughly 10-fold increase for about six months (Langbein, 1985). Also from two-color laser ranging, in the vicinity of the Long Valley Caldera, California, Linker, et al. (1986) report that deformation rates possibly related to an episode of magmatic intrusion decreased by a factor of 2 to 3 in 1983 - 1984. At longer time scales there is evidence that stress recovery following large earthquakes may be significantly nonlinear, apparently because of deep, time-dependent response to the strain step of the previous earthquake (Thatcher, 1983; Thatcher and Rundle, 1981).
Additional examples of stress events, some with correlated earthquakes are presented in this volume. It is concluded that non-linear stressing commonly occurs in earthquake prone areas. This paper presents a method for calculation of the conditional probability for recurrence of specific earthquakes under conditions of nonlinear fault stressing, including stress events. The effect of nonlinear stressing is investigated assuming either instantaneous onset of earthquake slip at a stress threshold or time-dependent failure.

PROBABILITY CALCULATION

Recurrence Independent of Stressing Rates. Calculations of the probabilities for the repeat of specific large, plate rupturing earthquakes have been presented for several regions and potential earthquakes (Lindh, 1983; Sykes and Nishenko, 1984). The method used in those studies requires an estimate of the expected recurrence interval for the earthquake, tr, the time since the last such earthquake, t, and an estimate of the probability distribution for the recurrence time. Expected recurrence intervals have been obtained from records of historical earthquakes, dating of geologic structures identified to have been formed at the times of previous events and by calculation from fault slip data. The calculations of expected recurrence intervals are based upon measurement of fault slip at the time of the previous earthquakes and the long-term average rate of accumulation of a slip deficit as measured by long-term fault slip rates or strain observations.
At present there is little information on the most appropriate function to employ for representation of the probability density for recurrence time. Lindh, (1983) and Sykes and Nishenko, (1984) use a Gaussian distribution. Buland and Nishenko (1986) propose that a log-normal distribution best fits the available recurrence data. Because the density distributions for recurrence are very broad,

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the probabilities obtained for earthquake recurrence are not especially sensitive to the function chosen, except at the extremes of the distributions. Following Lindh, (1983) and Sykes and Nishenko, (1984) we 'employ the Gaussian distribution for the current study. However, the approach presented here is not dependent upon the assumption of a Gaussian and other distribution functions such as the log-normal distribution could be substituted.
The probability that an the earthquake will occur in some interval, P(t, t+At), is found by integrating the the probability density function over that time interval:

-t+At

P(t,t+At) =jFdt (1)

Where F is the probability density function (i.e. the earthquake frequency as a function of time). For the case of a Gaussian distribution of recurrence time, tr, with a standard deviation, st:

F = (1/3 V5T)exp[-(t-tr) 2 /2s2] (2)

The probability for earthquake recurrence in the interval, t to t+At, conditional on the event having not occurred prior to t is given by the probability for the occurrence in that time interval divided by the probability for occurrence all times greater than t:

P(t, t+At)
CP(t, t+At) = (3)

pa too)

Recurrence with Variable Stressing to a Stress Threshold. The above formulation for earthquake probability is generally justified by a physical model of the earthquake cycle in which the earthquake occurs when the fault stress reaches a critical state. Although the model suggests that variations in stressing history are of importance in determining the time of the next earthquake, this formulation makes no provision for including observations of variations in loading during the earthquake cycle. For the following, we adopt the model for earthquake recurrence based upon elastic recovery of stress to some critical state (Fig. 1). The goal is to express the conditional probability problem in terms of stress state, which then permits incorporation of observations for variable stressing rates into the probability

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calculation.
As formulated here, all stress related effects are represented in terms of changes in the shear stress acting on the fault plane. The effect of normal stress variations in altering fault strength are not specifically considered here. However, such effects may be approximated by finding the equivalent change in shear stress, At, caused by the normal stress change, A(J, assuming Coulomb friction (i.e. Ax = -u. A<J). However, this approximation should be employed with caution because it ignores complications that arise from the dependence of fault strength on normal stress history as found by Linker and Dieterich (1986).
For convenience stress, X, is defined relative to the stress at the beginning of the cycle with X = 0 at t=0. Therefore, under the conditions of constant stressing rate, the stress, X, at any time in the cycle and the critical stress, X*, for the next earthquake are given by:

X = Rat, X*=Ratr (4)

Where Ra is the average rate of stressing during the earthquake cycle.
To express the probability problem in terms of stress state, we adopt the hypothesis that earthquakes occur at a stress threshold. Consequently, the probability density function for recurrence time may be expressed in terms of a probability distribution of the difference between the critical stress, X* as defined by (4) and the initial stress state. For ease of discussion, we represent this distribution entirely by the initial stres state and assign a unique value to X*. Hence, the standard deviation for stress, s^ , is defined by the standard deviation for the distribution of recurrence times, st, and the average loading rate:

sT =st Ra (5)

See Fig. 1. It is emphasized that the results are independent of this assumption. For example, identical results are obtained if the intial stress is taken to be uniquely defined and X* is given the probability distribution.
If the standard deviation of stress is held constant and the stressing rate, R, is not equal to the average rate, Ra, then the effective standard deviation of recurrence time, st', is no longer a constant, but varies with the instantaneous stress rate, R:

st' = stRa/R (6)

This relationship is illustrated in Fig. 2.

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For a Gaussian distribution, the probability density function for earthquake frequency at some stress state is:

F = (R /sTV2T) exp[-(T-T*)2 /2s*] (7)

V fc

The expression (1) for probability of an earthquake occurring in any interval t to t+At may therefore be written in terms of stress states at t and t+At:

rTt+At

P(t, t+ At) = (1 /sTV2T) exp[-(U-T*)2 /2s2.]di (8)

Where, Tt and Tf+At are me expected stresses at times t and t+At, respectively. Note that the R term in (7) arises from the definition of F as earthquake frequency (i.e. number of earthquakes/time) and that the term drops out for integration over T. Using the similar result for P(t, t^) obtained by integrating (8) over the interval (Tt, T^) the conditional probability as defined by (3) may then be determined. Expressed in this form, the conditional probability for earthquake slip at a critical stress threshold may be found for any loading history.
Note that the probability of triggered slip at the time of a stress step that increases the stress from
Tt to Tt+At, is simply:

____ |*W4- *-»

P (Tt, It+Al) = (1 /sTV27T) exp[-(U - T*)2 /2s2] dx (9)

Again, the conditional probability as defined by (3) is found using (9) and the probability of slip over the remainder of the distribution, (If., T^) as given by (8). The case of a stress step that results in a decrease of stress is addressed below.
Some effects of nonlinear stressing on earthquake recurrence can be seen by reference to Fig. 2. For any model of earthquake recurrence at a stress threshold, stress steps and changes in loading rate have the effect of altering the probability that the critical stress for the earthquake will be reached in some fixed time interval. Compared to the constant rate model, the width of the interval of the probability density function for earthquake stress that is sampled in a fixed time interval will be proportional to the stress rate. The probability rate and the increase in the probability with time will therefore scale

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accordingly with loading rate. Similarly, the portion of the probability density function for earthquake slip sampled in any time interval increases by the amount of any stress steps that occur in the period of interest.
In practice, it may be expected that the determination of conditional probability for the interval
(t, t+At) will usually employ the assumption that the loading rate at the time of the last observation will continue to act over the coming interval, At. A possible exception to this assumption would be in situations where there exists a specific model for nonlinear stressing. For linear stressing over the time interval of interest, the loading history is fully specified by the constant loading rate, R, beginning at some time, tQ, and the stress, TO, at time tQ. The probability density function for recurrence as a function of time may be then be found from (7) using (4), (5) and (6):

F = (R/Rast 2) exp{-[T0/Ra + (R/Ra)(t-tQ) - tr]/2st } (10)

The probability of earthquake recurrence for any interval t>tQ is obtained as before with (1) and (3). Therefore, to implement a calculation of conditional probability for variable stressing, values for the parameters tr, st, Ra, TQ and R are needed. Compared to the usual calculation of conditional probability using (2) the necessary new parameters are TQ, Ra and R. Given sufficient strain monitoring baseline data for the region of interest, it appears that it may be feasible to estimate Ra and the relative loading rate, R/Ra- Additionally, Ra is defined by tf and earthquake stress drop. It is expected that determination of 1Q will be substantially more difficult because it requires detailed knowledge of the stressing of the region for the entire interval, up to time t. As a practical matter this would seem unlikely and therefore, in the following, we assume constant stressing at Ra up to time, tQ, to obtain an estimate of TQ. Equation (10) may then be written:

F = (r/stV2nT) exp{-[q + rt -t,.] 2/2st2} (11)

Where:

q = t0 [l-(R/Ra)], r=(R/Ra) (12)

Note that under conditions of constant stressing rate during the entire cycle, q = 0 and r = 1 and (11) reduces to the original formulation based on (2) as used by Lindh (1983) and Sykes and Nishenko (1984). For the case of constant stressing at the average rate up to time, tQ, followed by a stress step, AT = and stressing at rate R:

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q = Ktr + t0 [l-(R/Ra)L r=(R/Ra) (13)

If the stress step is negative, that is the shear stress decreases from T + AT to 1, then the stress interval (T, 1 + AT) of the probability distribution must be traversed a second time as the stress subsequently increases. Because the earthquake did not occur on the first pass through the interval (T, T + AT) prior to the step, it cannot occur on the second pass under the assumptions of the model. Hence, the probability of earthquake recurrence in the second pass through any any interval (T, T + AT) must be zero. The duration of the time of zero probability, tq, following a decrease in stress is:

tq = AT/R (14)

Recurrence with Time-Dependent failure. Characteristics of earthquake occurrence and laboratory observations of rock friction and rock fracture indicate that the onset of earthquake faulting may involve rate-dependent properties not represented by the assumption of instantaneous earthquake failure at a constant stress threshold. Rate dependence of the critical stress for earthquake faulting will introduce additional time-dependencies of the earthquake probability not encountered under conditions of constant stressing rates or under the assumption of a constant stress threshold with variable stressing. Features of seismicity that have been proposed to be linked to time- or rate-dependent failure mechanisms include the general absence of correlations between earthquakes and the stressing from solid earth tides (Dieterich, 1987) and the occurrence of foreshocks and aftershocks (Das and Scholz, 1982; Dieterich, 1986). Laboratory observations have demonstrated significant loading rate dependencies of critical stress for both the fracture of initially intact rock (e.g. Carter, et al., 1981) and for slip on pre-existing faults (e.g. Dieterich, 1981).
In the following we apply failure criteria predicted by a rate- and state-dependent constitutive law for fault slip to the earthquake probability problem. Constitutive formulations of this type have been found to be an effective means for representing rate and rate-history characteristics of fault strength measured in the laboratory and for modeling of various sliding phenomena. Details of the the rate- and state-variable formulation, laboratory observations and applications to earthquake problems are given by Dieterich (1979a,b; 1981; 1986; 1987), Ruina (1980,1983), Rice and Ruina (1983), Rice (1983), Gu et al. (1984), Okubo and Dieterich (1984, 1986), Weeks and Tullis (1985), Tse and Rice (1986), Blanpied and Tullis (1986) and Tullis and Weeks (1986).
The approach to probability calculation presented here is based upon results for earthquake nucleation models with rate- and state-dependent friction (Dieterich, 1986, 1987a, 1987b). Additional details on the characteristics of the constitutive law and applications to the nucleation of unstable

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fault slip are given in those studies. The constitutive law for the coefficient of fault friction, p., is:

(15)

Where; M.Q, B and A are constants; 8 is the sliding speed (per second) divided by the characteristic slip distance, DC; and 0 is the state variable. For 0 in units of seconds and 8 in units of DC/S, experimental values of B and A typically fall in the range 0.006 - 0.010 and 0.003 - 0.007, respectively. For the results given below, the constitutive parameters appear as the ratio B/A. Values for B/A usually fall in the range from 1.2 to 2.2. The state variable 0 depends upon slip history. For this study we employ the following evolution law for 0:

d0/dt=l-08 (16)

See Dieterich(1986) for a discussion of (16). A feature of this particular evolution law, exploited below, is that it is amenable to derivation of closed-form solutions.
The limit to the 8-dependence of u. at high slip speeds determines the maximum possible friction and the onset of unstable slip. We define the onset of instability as occurring when the normalized slip speed reaches the value 8^. Consequently, the friction at instability:

At steady state slip, d0 /dt = 0 and therefore 0 = 1/8. From (15) the steady state friction becomes:

(18)

If [i * ng a consequence of (16) is that the state variable, 0, does not remain constant, but evolves with time and displacement. As a result jij and u.g are not constants, but depend upon the sliding history. At any applied stress in excess of u.g the state variable 0 will decrease with displacement and the slip speed may accelerate to 8j resulting in instability. The stress interval from T = p.sO to T = UjCJ defines the stress range for self-driven accelerating slip to instability. Consequently, this stress range may be thought of as a range of critical stresses for instability.
It has been shown in the earlier studies that unstable slip may initiate on a slipping fault patch only if the patch exceeds some minimum critical dimension. The critical patch size is a function of normal stress, loading conditions and constitutive parameters which include DC. Over a wide range of conditions, the logarithm of the time to instability increases approximately linearly as the difference

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(I /<J - jis) decreases.
The use of state-variable fault friction introduces an important complication to formulations for estimating conditional probability of earthquake recurrence. The stress at the instant of instability is not a constant. The stress at the onset of instability depends upon 0 as given by (17). In turn, 0 varies according to the details of the loading history and the initial stress. As a result of these dependencies, the probability distribution for recurrence time is no longer simply related to the stress distribution alone. Two methods of calculation of conditional probability have been employed in this study. The first avoids these complications by formulating the probability problem in terms of the initial stress state. The distribution of recurrence times resulting from the initial stress distribution and the loading history is obtained from a series of numerical calculations. Points in the probability distribution for recurrence time are found by beginning the calculation at an initial stress and marching through stressing history until the earthquake occurs. The second method is an approximate procedure that assumes that only the recent loading history is known. The second procedure permits the use of analytic solutions that simplify and and greatly shorten the computation.
The probability determination for the full computation employing the details of the loading history is straight-forward. In the following, the symbol, S, is used to denote the value of a specific initial stress within the probability distribution defined by the standard deviation, s^, and mean initial stress Xm. The goal is to find the initial stresses St and St+/^ t that result in earthquake instability at the beginning and end of the time interval of interest, respectively. Once St and S are found, the probability that the earthquake will occur in any interval, t to t+At, is given by:

f S t+At
P(t, t+ At) = (1 /s V2*T) expRS-V2 /2s2] dS (19)
Jc.

As before, the conditional probability is then obtained from (3) using the similar result for P(t, t^). The initial stress, St, resulting in failure at time t is calculated using the numerical model of Dieterich (1986,1987a). The model is for slip on a fault patch with dimensions that exceed the critical patch size for unstable slip. Stress - displacement relations for the center of the patch are represented through an equivalent patch stiffness that is obtained from the usual dislocation formulas. Comparisons with detailed plane strain and anti-plane strain models of the nucleation process have shown that this simple patch model yields results for the time of instability that are in very close agreement with the results from the more detailed calculations (Dieterich, 1987b). Because the time to failure and other details of the nucleation process are not sensitive to patch dimensions in excess of the critical dimension (Dieterich, 1986), an infinite patch radius was arbitrarily used for the calculations

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presented below. Beginning at the initial stress state, the calculations march through time to follow the evolution of slip up to the instant of slip instability. For the probability calculation, a predictor-corrector scheme, iterates on the calculations to find the specific initial stress that yields instability at the desired times, t and t+At. As shown below, for constant rates of stressing, the model with time-dependent failure yields probabilities that are nearly identical to those obtained from the simpler recurrence-time model. Small difference arise because of the slow increase in strength with time in the time-dependent failure model.
The approximate procedure for calculation of conditional probabilities is based upon a recently obtained set of solutions for slip on a fault patch with state- and rate-dependent friction (Dieterich, 1987b). Closed-form solutions have been obtained for slip as a function of time and patch radius for a variety of loading conditions. The loading conditions include stress steps, jumps in stressing rate, and sinusoidal varying stresses superimposed on uniform loading rates. Applied to populations of patches those solutions have been used to find the effect of stressing perturbations on seismicity rates (i.e. the probability density function, F). The solutions pertain only to the interval of accelerating slip prior to the instability, not the entire loading cycle and assume a constant probability density for earthquake frequency in the absence of a loading perturbation. The frequency, F ', following a loading rate change from R to R:

-i -1

(20)

And for a stress step:

F ' = F{1 + [h-llexpt-gtt-tpir1 (21)

In (20) and (21), F is the frequency immediately before the loading perturbation in the absence of time-dependent failure, tj is the time of the loading perturbation and:

F

l- ~
FR2 1 + _ l
exp[g(t-t)R2 /R1]

g = Rj/GA = ATgR-L/R
(22)
h = exp(-KAT e /A) = (8itrrKB/A

As defined for equation (13) the stress step AT is defined as a factor K of the earthquake stress drop (i.e. AT = K ATe). The above approximations of g and h using recurrence time, tf, follow from the

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dependence of the stress drop, Alg, of the previous earthquake on 0 at the time of that earthquake. The value of 0 at the time of the previous stress drop is approximated in (22) as the average recurrence time, tj.. This approximation represents an upper limit for 0.
Before addressing the question of recurrence probabilities, discussion of the application the state-variable constitutive approach to earthquake occurrence may now be offered based upon (20) and (21). Dieterich (1987b) applies these results to populations of nucleation patches to model seismicity rates for various load perturbations. A practical test of the appropriateness of the model for time-dependent failure is in application of (21) to aftershock occurrence. The application of (21) to aftershocks is based on the assumption that the stress redistribution caused by the mainshock in turn causes aftershocks. Evidence supporting this assumption include: i) the characteristic clustering of aftershocks at the edges of the mainshock rupture, where the stress increase due to earthquake faulting is greatest, ii) the occasional occurrence of lesser clusters of aftershocks not on the primary fault for the mainshock but, in other areas where dislocation analyses show the stress to have increased due to the mainshock faulting (e.g. Stein and Lisowski, 1983). iii) Aftershock-like jumps in seismicity rates followed by characteristic decay in rates with time for non-earthquake related jumps in stress. A notable example of this is the perturbation of background seismicity following dike intrusion episodes at Kilauea volcano (Dvorak and Tanigawa, 1985).
The result (21) is worth emphasizing because it yields predictions of aftershock rates following the stress step of a main shock, that can be compared to the well-known Omori's Law for aftershock decay:

F'= Fv/fw + a-tj)] (23)

In (23) the aftershock rate predicted by Omori's Law is normalized by the background seismicity rate, F, prior to the mainshock. As shown by Fig. 3, (21) and (23) yield essentially indistinguishable results. The principal difference is that Omori's law (23) unlike (21) makes no provision for return of seismicity rates to the background frequency F. To the extent that (23) is an adequate general description of aftershock rates, the agreement of the theory (21) provides a necessary, but not sufficient confirmation of the approach presented here. Additionally, the prediction of Omori's Law by (21) opens an opportunity to use aftershocks decay parameters to determine physical model parameters. For example, the normalized earthquake rate immediately following a stress step [at (t-tp=0] is:

F 7F = v/w =l/h = ((24)

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The duration of aftershocks, t_:

ta = v-w = 1/g = RaA tr /(R1B In 8^ (25)

Hence, the rate of aftershock seismicity is determined by both the recurrence time and the magnitude of the stress step; and the duration of an aftershock sequence is determined by the recurrence time alone. Both aftershock duration and aftershock rate depend additionally on the ratio of constitutive parameters B/A. This particular application of (21) to aftershocks ignores a number number of expected complications. Two topics of interest that are under current investigation: i) interactions due to stress redistributions within the aftershock sequence (i.e. aftershocks to the aftershocks) and ii) spatial variations of the magnitude of the stress step and apparent speading of aftershocks with time.
To employ (20) or (21) for recurrence of a specific earthquake subsequent to a loading perturbation it
is necessary to use an unperturbed probability density, F, which is appropriate to the recurrence of a single earthquake. The derivation of (20) and (21), is for constant F, while for earthquake recurrence, F has a stress or time-dependence as given previously (e.g. equation 11). In the following, the assumption of constant F has been ignored and the time-dependence of F is treated by simply substituting (11) into
(20) or (21). It is noted that the probability density function for recurrence times of (2) is very broad (i.e. st is typically 0.3tr) while, the duration of the perturbation of F by time-dependent failure is relatively short (i.e. ta is typically on the order of 0.03tj.). Hence, during the time that F is perturbed by the time-dependent failure effects, F will not change significantly. Consequently, the substitution of
(11) for F is not expected to introduce serious error. Comparison results from the full numerical calculation with this approximation, presented below, show good agreement and bear out this expectation. Specifically, for earthquake recurrence following a change of loading rate:

F 1 =

Ri >(q /^i /2s) R
R, R

- 1

exp[g (t-t) R2

- 1

(26)

and for a stress step:

R exp{-(q + Rt/R -t)/2sf } F ' = i i a J- L_ x + [h-l]exp[-g(t-t)]} (27)

In (26) note that the value of r which appears outside of the exponential in (11) has been set equal to

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Rj/Ra and the value of r appearing in the exponential has been set equal to R2/Ra- These substitutions satisfy both the requirement that F at the time of the rate step be equal to the frequency in the absence of a step (which scales by Rj) and the requirement that after the rate step F changes at the rate determined by the new loading rate R2>
The probability for recurrence of a specific earthquake is then obtained in the usual way by integrating (26) and (27) over the interval of interest:

t+At

P(t, t+ At)

-

= F ' dt
J

(28)

Solution of (28) for P(t, t+At) and P(t, t^) to obtain CP(t, t+At) may be accomplished by numerical integration. However, because of the very different time scales that characterize F and the perturbation of F due to time-dependent failure, more direct solution of (28) is possible for most situations of interest. If (t-tp or At are large (i.e.. much greater that ta), the perturbing effect of time-dependent failure becomes negligible and (26) and (27) approach (11). Solutions for (11) may be obtained from standard math tables or by numerical integration.
For short time intervals with At on the order of days to weeks we may assume that F is sensibly a constant and move those terms outside of the integral and explicitly integrate (28). For a loading rate change from R^ to R2 this yields:

P(t, t+ At) =

+ R 2t/Ra -ti)2/2st)
In

)gR2 /R1H+R2/Ri
exp[(t -t1)gR2 /R1]-l+R2 /R1

(29)

For a step in stress:

P(t,t+At) -

exp[(t+ At-Og]+h-l
exp[(t-tj)g] + h-l

(30)

In (29) and (30), q is evaluated in the absence of the rate change or stress step, respectively. For the results presented below, the conditional probability CP(t, t+At) is obtained using P(t, t+At) as given by
(29) and (30) and P(t, t^) as given by numerical integration of (11).

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From Dieterich (1987b) the probability of triggering earthquake slip at the time of the stress
step is similarly obtained:

P (lf, If+Al) = In
g

g/h (31)

Equation (31) is obtained from explicit integration and assumes F is constant over the stress interval. As with the case of instability at a constant stress threshold, a stress step that reduces stress on
the fault, will result in some portion of the probability density function being traversed a second time with zero probability of earthquake recurrence. The duration of the interval of zero probability following a decrease of stress is found to be (Dieterich, 1987b):

(32) 1 ,

t q - In
4 g

g/h + 8; B/A

g + Si B/A

Following the quiescence of duration tq (i.e. [t-tj] > tq) the probability of recurrence returns to that obtained from (30).

RESULTS

The following examples were computed to examine some effects of non-linear stressing with models for earthquake recurrence at a threshold stress and for time-dependent failure. The results for recurrence at a stress threshold were obtained by numerical integration of the probability density function (11). The results for time-dependent failure were obtained either by numerical integration of
(19) using initial stresses obtained by the time marching calculations for earthquake nucleation or by the explicit approximate relationships (29) and (30). Each of the examples assumes recurrence of an earthquake, roughly approximating the conditions at Parkfield, California, with tr = 22yr, st = 0.3tr/ Ra = Ale/22yr and 8^=0.3 Ale where the previous earthquake occurred in 1966 (Lindh, 1983; Sykes and Nishenko, 1984). For the time-dependent failure model, the constitutive parameters B and A are 0.008 and 0.004 (6 and 8 in s and DC /S, respectively) yielding a ratio of constitutive parameters, B/A, of 2.0 . The critical slip speed for unstable fault slip, 8j, is arbitrarily set at 1.0 DC/S.

Stressing Rate Effects. Examples illustrating effects of loading rate steps on probability for

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earthquake recurrence with failure at a fixed stress threshold are given by Fig. 4. The examples employ a stress state at t=17 years obtained through an assumption of loading at rate Ra up to that time. Beginning at year 17 in the cycle different stress histories are imposed. The stressing rates for those examples are in the range of Ra ± 0.5Ra. This range of loading rates is well below the changes in strain rate that have been described for some tectonic areas. The probabilities are for recurrence in a one year interval, conditional on the earthquake not having already occurred. The heavy curve (b) corresponds to continued loading at rate Ra and is the result obtained with the probability density function for recurrence time (2). For the example with alternating strain rates, the computation of the conditional probability assumes that the instantaneous stressing rate at the time, t, will continue to act throughout the time interval, At. Note the direct scaling of probability by the instantaneous loading rate. The jump in probability at the time of a stressing rate step arises from the multiplication of the probability density by the factor r = R/Ra which appears in equation (11). The scaling of the slope of the probability vs. time curves by the loading rate similarly arises from the multiplication of elapsed time, t, by the factor r in (11). Because of this dependence on loading rate, the probabilities for the example with alternating stress rates undergo large jumps even though the stresses depart only slightly from the average stressing curve. Those variations in probability due to the loading rate variations exceed the time-dependent changes in probability that occur at constant stressing rates over time periods equal to

±st.

Figures 5, 6, 7 and 8 show the effects of stressing rates on earthquake probability using the rate-
and state-dependent constitutive model. The examples again assume earthquake recurrence parameters as employed by Lindh (1983) for Parkfield, California. The probabilities obtained using the fixed threshold stress model are also plotted. In Fig. 6 at constant stressing rate, note that the probabilities for time-dependent failure are nearly identical the probabilities for the threshold stress model and the recurrence model of Lindh (1983) and Sykes and Nishenko (1984). The slight deviation of the time-dependent model arises from the strengthening of the fault with the logarithm of elapsed time. Note also in Fig.6 that the approximate explicit equation for probability is in good agreement with the numerical calculation. Compared to the model for failure at a constant stress threshold the addition of rate- and state-dependent properties to the earthquake recurrence model introduces additional time-dependencies to earthquake probabilities. The jumps in probability that occur at the time of loading rate changes tend to be attenuated and the probabilities subsequent to the steps asymptotically approach the probabilities obtained from the threshold stress model.
Because the probability density function for time-dependent failure asymptotically approaches
the threshold stress case, the probability determinations for long time intervals tend to be similar to those for the threshold stress model. Similarly, for the probabilities for small time intervals, at large elapsed times from the rate step, approach the threshold stress probabilities. For time intervals of one

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year (Fig.5) large steps in probability occur at the time of the stressing rate change. However, these steps in probability arise from averaging, over the entire interval, of changes in the short-term probability late in the time interval and not at the beginning. This can be seen in the results for the probability of failure in 5-day intervals (fig.6) which do no show the large jump in probability seen for the one-year interval. These effects are also evident from examination of the explicit equation for probability density following a stressing rate step (26). Immediately following the rate step, the probability density from (26) is identical to the probability density obtained for the threshold model
(11) immediately before the step.
Probabilities for the loading conditions consisting of stressing rates that alternate between 0.5Ra
and l-5Ra have been computed for the model with rate- and state-dependent fault friction (Fig. 7 and 8). As seen in the previous examples, the introduction of time-dependent failure results in a reduction of the jumps in probability that occur in the threshold stress. The effect is especially evident in for 5 day intervals (Fig.6). For time intervals of 1 year (Fig. 5) the differences between the threshold stress model and the time-dependent failure model are small.

Effect of Stress Steps. In nature, stress steps, mostly of small amplitude, can be expected to be a characteristic feature of the loading cycle because of the occurrence of nearby earthquakes. Adjacent to earthquake faulted segments, the amplitude of the stress step may be arbitrarily large because of the stress concentration at the ends of the rupture. If the step is sufficiently great, slip will occur at the time of the step and the earthquake will expand to include that segment.
The examples in Fig.9 give probabilities for recurrence following steps in stress for the threshold stress model. The principal characteristic here is a jump in probability that scales by the magnitude of the stress step. A stress step, AT = KAle, simply offsets the probability curve in time by an amount proportional to the loading step. The increase in probability at the time of the step is equivalent to the increase in probability occurring over the time interval KAle/R in the absence of a stress step (i.e. the recurrence time has been advanced to tr(l-K) as given by (11)). Because the time-dependency of probability under constant loading rate is relatively weak, the increases in probability from stress steps are also relatively small, unless the stress steps are very large.
In contrast to the threshold stress model, recurrence probabilities following stress steps obtained with the model employing the rate- and state-variable fault friction show large, short-term jumps. Fig. 10 gives results for the conditional probability for earthquake recurrence computed for 5 day intervals following steps in stress. The loading rate before and after the step is at the average rate, Ra. Note the good agreement of the approximate equation (30) with the numerical results. Prior to the stress step, the 5-day conditional probability is 0.0016. Immediately following a step of K = 0.10 the 5-day conditional probability jumps by a factor of 37 to 0.06 then rapidly decays and asymptotically

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approaches the offset given by the threshold stress model. For comparison, with the threshold stress model for failure the probability jumps only by a factor of 1.4 under this circumstance.
Probabilities of triggering an earthquake at the time of a stress step are illustrated in Fig. 11.
The examples are for steps applied at year 22 in the cycle with loading rates before and after the step at the average rate, Ra. The probabilities were computed from (9) and (31) for the threshold stress and time-dependent failure models respectively. Of particular interest, note that the probability of instantaneous triggering of the earthquake is very much larger for the threshold stress model. With the threshold stress model an interval of the probability density distribution equal to the stress step will pass through the failure stress at the time of the step. If the actual stress falls in that interval, then the earthquake occurs instantaneously at the time of the step. The lower probability of triggering an earthquake at the time of a stress step in the time-dependent failure model arises because there exists a finite range of stresses over which earthquake instability can nucleate. Except for a narrow interval of the probability distribution which results in instantaneous failure, the remainder of the interval yielding instantaneous instability in the threshold model is for recurrence at some time following the step.
The gain in probability of recurrence at the time of a stress step is:

[P(t, t+At)/P(t,
GAIN = (33)

where, the quantities in brackets labeled before and after are the conditional probabilities before and after the stress step, respectively. Equation (33) may be evaluated explicity for the case of small time intervals and time-dependent failure. The difference between P(t, totJbefore anc* ^*' ^^ after * s t^ie probability of triggered slip at the time of the stress step. Hence, the P(t, t^) terms cancel because the probability of triggered slip is negligible for all but very large stress steps with the time-dependent failure model (see Fig. 11). For short time intervals (i.e. small probabilities) P(t, t+At)^ore is simply FAt. Therefore substituting from (10) to evaluate the probability density, F, before the step and using
(30) to evaluate P(t, t+At)fl #er , the probability gain for time-dependent failure for short time intervals:

GAIN= in

gAt

exp(gAt) + h - 1
(34)

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Figure 12 plots the probability gain from (34) for various recurrence times. Note from Fig.12 and equation (34) that the probability gain is independent of the time since the last earthquake. The probability following a stress step is the GAIN multiplied by the probability in the absence of the step. Although the probability gain for a given stress step increases with tr, the conditional probability is less for longer tr Fig.12 shows that for any magnitude stress step, the probability gain is larger for longer recurrence times.
Finally, Fig.13, gives the probabilities immediately following a stress step for different time intervals, assuming a mean recurrence time of 22 years. The results were obtained from (30), but could also have been obtained by multiplying the pre-step probability from integration of the density function (11) times GAIN obtained from (34). The figure illustrates the relationship between the magnitude of a stress step, probability of earthquake recurrence and time following the step. For large stress steps the probability of slip within a few seconds or minutes of the step become significant. For an earthquake induced stress step these results suggest a continuous relationship between triggered slip at the time of the stress step, composite earthquakes events occurring within a few seconds of one another, and paired earthquakes and mainshocks following foreshocks occurring within minutes to days.

DISCUSSION AND CONCLUSIONS

Ishibashi (1987) proposes that earthquake precursors may be divided into two broad classes: physical precursors and tectonic precursors. Physical precursors are symptomatic of the failure process for an impending earthquake, while tectonic precursors may be viewed as symptoms of regional stress/strain changes in which the occurrence of earthquakes is but another symptom. The tectonic precursor may therefore be thought of as a stressing change that increases the probability of earthquake occurrence. Examples noted in the beginning of this paper and other examples described by papers in this volume suggest that stressing variations in tectonic regions may commonly occur.
This paper proposes a possible extension of the method employed by Lindh (1983) and Sykes and Nishenko (1984) to calculate the probability of earthquake recurrence under conditions of variable stressing. It has been shown that variations in fault stressing result in significant variations of the probability of earthquake recurrence when specific models for the mechanics of earthquake slip are incorporated into the recurrence formulation. Two specific models for the mechanics of earthquake slip have been considered here. The first is for earthquake slip at a constant critical stress threshold for unstable fault slip. The second is based upon laboratory measurement of fault properties that demonstrate a dependency of the stress at the time of unstable slip on stressing rates and previous

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stressing history.
In the context of the threshold stress model, the calculation of recurrence probability by Lindh (1983) and Sykes and Nishenko (1984) represents the simplest approximation for stress-dependent recurrence where the stressing rate is always at the average rate Ra. The consequences of adding stress-dependence to the original probability density function for recurrence time may be summarized as follows: When the stressing rate, R, is not equal to the average rate, Ra, the probability density function (11) scales by the multiplying factor r=R/Ra. Therefore, the probability for recurrence at any time in the cycle will also scale directly by r. Under conditions of constant stressing the probability of recurrence increases with time because the probability density is a function of, t, the time since the previous earthquake. The effect of changing the stressing rate, R, is equivalent to running the clock for elapsed time, t, faster or slower by the factor r in equation (11). Hence the slope of the curve for probability vs time scales directly by r. The parameter, q, appearing in (11) is the amount by which the elapsed time clock is set ahead or behind at the time to in the recurrence cycle. The effect of a stress step is to set the clock for elapsed time ahead by the factor Ktr as given in (13).
Although, the threshold stress model has intuitive appeal and yields relatively simple formulations for recurrence probabilities it is not considered appropriate for general application. This conclusion is based on the failure of the model to account for various time- and rate-dependent fault properties observed in laboratory experiments and inferred for the earth. These time- and rate-dependent fault properties introduce additional rate- and time-dependent effects into the probability determinations. Specific evidence supporting the conclusion on the inadequacy of the threshold stress model is the inability of that model to predict higher probabilities for an interval following steps in applied stress, as apparently required by the occurrence of foreshocks and paired earthquakes.
The time-dependent failure model proposed here overcomes the shortcomings of the threshold stress model. The time-dependent failure model is derived from a rate- and state-dependent constitutive law for faults. This constitutive law is based upon experimental observations of fault slip and has proved to be effective for modeling a variety of phenomena observed in laboratory experiments for modeling of different faulting and earthquake processes. The successful prediction of the aftershock decay law following a stress step is a specific test of the time-dependent failure formulation proposed here. This subject is discussed below.
The principal effect of adding the time-dependent failure criteria is to introduce transient variations in probability following changes in stressing rate or stress steps. Following a change in stressing, the probabilities for time-dependent failure asymptotically approach the probabilities obtained for the threshold stress model. Consequently, the effects of time-dependent failure are most evident in intermediate- and short-term probabilities following stressing perturbations. For

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sufficiently long time intervals at constant stressing rates, the threshold stress model is completely adequate.
The effect of loading variations may be illustrated further by a hypothetical example. Using Lindh's parameters for the Mojave section of the San Andreas fault (tr=145yr, st=44yr) the constant straining assumption yields an annual probability of 0.013 for 1986. If the earthquake has not occurred prior to the expected recurrence time in the year 2002 the probability will then have increased to 0.018. Using the stress threshold model with stress rate equal to the expected long term average stressing rate, and a stress step in 1986 of K=0.11 the annual probability would jump immediately to the 2002 value of 0.018, a gain by a factor of 1.4. The same stress step with the time-dependent failure model is yields a probability of 0.019 for the 5 day interval following the step, which is an increase in the probability by a factor of 107. An increase of the loading rate to 8 times the mean rate as possibly observed in 1984 (Langbein, 1985) would result in a probability of 0.11 per year with the threshold model. The 8-fold increase in stressing rate with the time-dependent failure model yields an annual probability of 0.036 immediately following the rate change and increases to 0.052 and 0.070 after elapsed time of 3 and 6 months, respectively.
An interesting and possibly significant implication of this study is that stressing variations equally affect the probability of recurrence of a specific large earthquake of interest and the probability of the smaller earthquakes that comprise the background seismicity. Stressing variations alter the probability density of both the recurrence of a specific large earthquake and the population of smaller earthquakes in the region of interest. For a population of earthquakes, the probability density at any time is simply the observed earthquake rate. This means that the variation of the probability density for earthquake recurrence arising from stressing variations will be given directly by the observed variation of the rate of background seismicity in the nucleation zone of the expected large earthquake, assuming the background rate is constant in the absence of stressing perturbations.
Further study will be needed to establish the usefulness of this proposed relationship between background seismicity rate, stressing variations and probability rate for earthquake recurrence. However, it is argued that the predicted relationship has already been empirically established for the case of stress steps. It is well known that a stress step causes a 1/t decay of aftershock rates and it has been established for foreshock data that the rate of mainshock occurrence following a foreshock similarly decays by 1/t (Jones and Molnar, 1976; Jones, 1985). Additionally, Jones and Lindh (1987) present a formulation for conditional probability of recurrence at Parkfield, California following possible foreshocks that has the essential characteristics found in this study for probabilities following stress steps. The agreement of the result given by equation (19) for earthquake rates with aftershock rates and mainshock rates following foreshocks is considered to be significant. It is a specific test of the theory and therefore it provides a basis for concluding that method presented here adequately predicts

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the effects of more general stressing variations on earthquake rates and recurrence probabilities.
In conclusion, it is noted that the increase in probability with time as obtained with the stress-independent recurrence model of Lindh (1983) and Sykes and Nishenko (1984) is relatively small because the probability distributions for recurrence time are quite broad. These small variations of probability are in contrast, to the large variations in probability that arise from even modest stressing variations with the models presented here. The stress-independent recurrence probabilities have been employed to evaluate the potential for earthquake recurrence of various fault segments. This study shows that stressing variations, when they occur, can be more significant that the simple progression of time in determining the probability of recurrence on a specific fault segment. Consequently, observations of stressing rates and seismicity rates may prove to be of practical use for monitoring changes in recurrence probability especially for intermediate and short time scales.

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Fig. 2 Model of earthquake cycle with variable stressing. The average stressing rate is Ra The standard deviation of stress, s_ , is constant and the resulting standard deviation of expected recurrence time, s J varies from s when R does not equal Ra.

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K = stress step / earthquake stress drop

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Fig. 3. Earthquake frequency following a stress step. Heavy line is theory, light line is Omori's Law for aftershock decay fit to the K = 0.35 result. Constitutive parameters A and B are 0.004 and 0.008, respectively. Aftershock duration is t .

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Fig. 4. Conditional probability per one year interval for earthquake recurrence, assuming slip occurs at a constant stress threshold. Average recurrence time is 22 years and standard deviation is 6.6 years. Beginning at year 17, the loading rates step to 0.5,1.0 and 1.5 times the average rate for a, b and c, respectively. The dashed curve, d, alternates between 0.5 and 1.5.

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Fig. 5. Conditional probability for earthquake recurrence per one year interval as calculated with the numerical model for rate- and state-dependent fault friction (heavy lines). The model assumes average recurrence time of 22 year and a standard deviation of 6.6 year. Beginning at year 22 the loading rates step to 0.5,1.0,1.5 and 2.0 times the average loading rate for curves a, b, c and d, respectively. The constitutive parameters A, and B have values of 0.004 and 0.008, respectively. The light lines are for slip a fixed stress threshold.

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Fig. 7. Conditional probability for earthquake recurrence per one year interval. Beginning at year 17 the stressing rates alternate between 0.5 and 1.5 times the average rate as in Fig. 4. Average recurrence time is 22 years with a standard deviation of 6.6 years. The heavy curve employs rate- and state-dependent fault friction, as calculated with the numerical model. Constitutive parameters A and B are 0.004 and 0.008 respectively. The light curve is the result of Fig. 4 which assumes failure at a constant stress threshold.

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TIME (YR)

24 25 26 27

Fig. 8. Conditional probability for earthquake recurrence per five day interval. Beginning at year 17 the stressing rates alternate between 0.5 and 1.5 times the average rate as in Fig. 4. Average recurrence time is 22 years with a standard deviation of 6.6 years. The heavy curve employs rate- and state-dependent fault friction, as calculated with the numerical model. Constitutive parameters A and B are 0.004 and 0.008 respectively. The light curve is the assumes failure at a constant stress threshold.

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K = stress step / earthquake stress drop

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Fig. 9. Conditional probability for earthquake recurrence per one year interval for slip at a constant stress threshold. Expected recurrence time is 22 years with a standard deviation of 6.6 years. The loading rate is at the average rate before and after the stress step K.

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Fig. 13. Conditional probability for earthquake recurrence immediately following stress step K at year 22. Average recurrence time is 22 years and the standard deviation is 6.6 years. The loading rate is at the average rate. The calculations employ equation (30) for the time intervals indicated on the curves.

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Summary of session on Statistical Methodology

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This session included seven papers, most of which discussed aspects
of analysis of seisraicity data. The seisraicity techniques discussed varied widely, from qualitative descriptions of recent activity in Southern California to quantitative statistical analysis of data from California and various other regions of the world. The non-seismicity papers presented general discussions of aspects of intermediate-term precursors and the effect of time-dependent failure on recurrence models.

Statistical methods for investigating quiescence and other temporal seismicity patterns by Mark Matthews and Paul Reasenberg.

In this presentation Matthews made several important points. First, he pointed out the necessity of quantifying observations of changes in seismicity patterns and decried the lack of such quantification in most past studies. He then described a new statistical approach to identification of periods of anomalous seismicity rates based on the study of stochastic point processes. An important difference between this approach and those that have been taken in the past is that the significance levels are determined through a set of simulations, rather than on the basis of some assumed distribution of the test statistic.

Precursory seismic quiescence: A critical assessment of the hypothesis by Paul Reasenberg and Mark Matthews.

Reasenberg presented results of the application of the statistical technique described by Matthews to data from: 1) Past mainshocks in California and Japan, 2) Characteristic earthquake sets from Bear Valley, California, and 3) Segments of the San Andreas studied by Wyss and Burford. His important conclusions were: 1) The nature of seismicity rates in the regions studied for the past mainshocks were highly variable. Some regions showed common excursions with 90% significance and others showed uniform rates. 2) No evidence was found for a systematic, widespread, or reliable pattern of quiescence prior to the mainshocks. 3) Small regions surrounding characteristic event hypocenters in Bear Valley showed uniform rates over an entire earthquake cycle. 4) The quiescence used by Wyss and Burford to successfully predict the May 1986 event was significant at the 97% level, but the others they reported were not significant using this test. 5) The presently occurring anomaly in the San Juan Bautista seismic gap is significant at the 71% level, and therefore, cannot be considered as an anomaly. 6) A significant seismic quiescence exists in the Parkfield, California region.
Reasenberg and Matthews proposed two numbers for evaluating various proposed precursory behaviors. These were the reliability (number of successful predictions / number of mainshocks), and the validity (number

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of successes / number of anomalies). From their study of past mainshocks they concluded that quiescence was not a reliable precursor. The rather lively discussion this paper centered on the need to incorporate knowledge of the tectonics of the region being examined into studies of earthquake prediction. Reasenberg and Matthews purposely avoided use of such knowledge. It was also pointed out that the beginning of the quiescence in the Parkfield data corresponded to the start of the preliminary catalog for that region.

Anomalous seismicity in the San Diego coastal region by Tom Heaton.

San Diego has always been a region of California where one can live
and not feel earthquakes. This folklore reflects the low rate of historical seismicity in this region. This low rate makes the recent upturn in seismic activity in this region stand out strongly. Heaton examined the rates of occurrence of events with ML > 3.5 prior to recent large events in California qualitatively. He found that the rate of activity increased prior to large earthquakes in the Mammoth, Coso, Imperial, and Coalinga regions and that no increased rates were obvious prior to the San Fernando or Borrego Mountain earthquakes.
Heaton then pointed out the tectonic similarities between the regions with increased activity prior to the larger earthquakes. With the exception of the Coalinga event, the increased activity prior to larger shocks occurred in regions of strike-slip faulting accompanied by crustal spreading. He pointed out that the region off shore from San Diego has physiography very similar to that in these regions, with an obvious series of basins and ranges.
On the basis of the qualitatively described seismicity patterns and
the tectonic similarity of off-shore San Diego and regions with possible precursory increases in seismicity Heaton concluded that we have not seen the last of the seismic activity in the San Diego region. Heaton also pointed out that Duncan Agnew had recently purchased earthquake insurance. This talk represented well many of the difficulties involved in real-time precursory evaluation.

Precursory seismic quiescence by Max Wyss and R. E. Habermann.

Wyss pointed out that our understanding of the phenomena of precursory quiescence was limited by the small number of well documented cases. He presented a summary of 17 cases of precursory quiescence which satisfied fairly stringent criteria and tried to characterize the phenomena on the basis of those observations. In the best cases the spatial and temporal extent of the anomaly were well known, the significance of the anomaly had been assessed, the catalog had been carefully studied to eliminate man-made causes for the quiescence, and the surrounding region had been searched for false alarms.
Wyss showed that in most of the good cases the backgound and anomalous rates were well defined and showed small variances. The anomalous rates were generally about 30% of the background rates, making many of the precursors resilient to errors in analysis or problems with the data. In four cases where the seismicity was high enough to allow sub-division of the rupture zone, the region close to the epicenter did

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not show precursory quiescence.
The relationship between precursor time and mainshock magnitude was examined for these 17 cases and clearly showed several interesting observations, though the amount of data was small. The data did not fit the relationship of Scholz et al., 1973, and the data from the creeping segment of the San Andreas fault showed precursor times that were clearly too long compared to those for events in other tectonic settings. Wyss speculated that this may reflect a difference in mechanism with precursory fault creep causing stress relaxation and quiescence in the volume around the fault in some cases and decreases in slip rate (creep retardations) causing quiescence on the fault in the creeping segment.

Plausibility criteria for intermediate-term earthquake prediction by Dave Jackson.

Three criteria were proposed for useful geophysical anomalies: 1) Anomalies must be observed simultaneously at several sites; 2) Coseismic anomalies must be observed at the same sites; and 3) The magnitude of the anomaly must bear a reasonable relationship to epicentral distance. Jackson pointed out that few of these criteria were met by any geodetic anomalies reported to date and addressed the reasons for this.
Low instrument density is a major factor in the lack of coherent precursors. The chief cause of this problem is lack of available resources. The lack of coseismic changes could be due to locations of instruments near nodes of the stress pattern due to the earthquake. "Sensitive spots" may explain the lack of reasonable amplitude vs. distance.
The solution to this problem outlined by Jackson involved a number
of steps. We need to support more basic research in measurement and provide suitable rewards for people making instruments. The goal of this work would be the development of inexpensive instruments which could be widely deployed at a reasonable cost. We need to be more careful about site selection and calibration of sites using rainfall effects and tides in order to take advantage of the small number of instruments we can afford. Last, and most importantly, we need to take a hard look at the costs of what we really need to solve the earthquake prediction problem and find a way to raise the necessary funds.

Some observations and some thoughts on intermediate-term precursors by Selwyn Sacks and Alan Linde.

A general description of problems involved in reliably recognizing intermediate-term strain precursors was presented. The central point of the discussion was the differentiation of noise and signal in situations where you had no idea what the noise looks like, or what the signal looks like. An example of two instruments which were 5 to eight km apart and saw very similar signals was shown to make the point that because these two instruments saw the same thing we would call it signal, but if the same observation was made on only one instrument, it would be noise. If this is so, how can we succeed in recognizing real precursors?

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Sacks pointed out that the main source of the noise in strain measurements was the sites and that the sensitivity of the instruments was more than adequate to observe expected signals. All sorts of things can cause noise at sites and if these noise sources (for instance local fractures) are concentrated near a site, it will be a bad site. Tidal amplitudes can be used to determine site sensitivity at short periods but the sensitivity at intermediate or long periods may be quite different.
Finally Sacks developed a case from the Izu peninsula in Japan. A large strain event occurred inside of a linear array of 10 borehole strainmeters. The largest strains observed were near the center of the network, so presumably the source of the disturbance was near that location. Only five of the nine possible sites which could have recorded the event did so. Sacks examined the drift rates at the sites and concluded that they were not a reliable indicator of whether the site recorded the event or not. He found that the best indicator of intermediate to long-term sensitivity was the amplitude of the annual cycle due to rainfall.

Probability of earthquake recurrence with non-uniform stress rates and time-dependent failure by Jim Dieterich.

Most methods of determining conditional probability of earthquake occurrence assume constant stress rate and failure at a given threshold stress. There is, however, plenty of evidence for variations in stress rates in the earth as well as for time dependent failure. Dieterich developed a model which considered the effects of stress (or strain) rate changes on the conditional probabilities.
In the case without time-dependent failure, the stress rate scales directly into the probability gain as a function of time. The dependence on loading rate means that even small changes in stress level, if they happen quickly, can cause large changes in probability. The addition of time-dependent failure into the model for stress rate changes smooths out the changes in probability and they approach the changes in the threshold stress model after several years. Dieterich concluded therefore, that stress rate changes did not alter the conditional probability substantially when time-dependence was added to the model.
The results for imposed stress steps were substantially different
for threshold and time-dependent models of failure. In this case the threshold model shows an upward offset in the conditional probability. The time-dependent model shows a sharp increase in probability (his example showed an increase in probability by a factor of 37 as compared to 1.4 for the threshold failure model). He concluded that, if time-dependent failure occurred in the earth one would expect much larger variances in recurrence times than those predicted by models which do not consider these factors.

Problems, New Results, and Future Work

The principal problem highlighted by this session is the lack of well developed case histories of intermediate-term precursors.

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Developing these case histories is possible in the case of seismicity patterns. We are beginning to overcome many of the problems in the catalogs and it will soon be possible to increase the number of carefully studied cases substantially. It is clear that this work can benefit greatly from increased understanding of tectonics and fault behavior. The papers presented by Matthews and Reasenberg showed that applying even a possibly excellent technique indiscriminately will lead to results which are difficult to interpret.
In the case of geodetic precursors, it seems that little can be done at present to increase the amount of precursory information which is presently available except hope for earthquakes or other strain events near instruments. It is clearly important to pick the sites as carefully as possible and to find a way to evaluate the sites reliably. Of course it would be great if we could do this without the expense of installing the instruments, but this seems unlikely.
A second point made clear in this session by Tom Heaton is the difficulties involved in dealing with the real-time occurrence of a seismicity pattern which is a possible precursor. This is similar to the noise-signal separation problem addressed by Sacks. The recent activity in the San Diego region is certainly interesting, but what do we do with it? The Aleutian Island case developed by Kisslinger and the Stone Canyon case of Wyss and Burford also showed that we are very inexperienced in dealing with earthquake predictions which are based on possible precursors rather than on some repeat time estimated from a recurrence model. Hopefully this will change with more high quality analysis of existing data and with the development of more case histories.
The final important point raised by Dieterich is also related to
the interaction between long- and intermediate-term precursors. His work serves as an important reminder that the long-term probabilities calculated from recurrence models are only general indicators of seismic risk. The actual probabilities may change substantially in real time under the influence of stress rate changes or stress steps. This is well demonstrated by the work that Simpson presented later in the conference concerning interactions between Parkfield and Coalinga. In that case it seems that the characteristic event is likely delayed by the Coalinga event, so the actual probabilities are lower than those expected solely on the basis of time predictability.
In any case, it seems clear that the earth is complex and that precursors are likely to be complex as well. Jackson was right when he said that if precursors are complex earthquake prediction is likely to be a very difficult task. I think he was also right when he said that we must be aware of the costs involved in doing it right and aggressively pursue the necessary funding.

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Rock Friction Constitutive Behavior from Laboratory Experiments and

Its Implications for an Earthquake Prediction Field Monitoring Program

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Abstract

Laboratory experiments by a number of investigators studying the constitutive properties
of frictional sliding of rocks have shown that the frictional resistance depends upon the velocity of sliding and has a memory of previous resistance that fades as a function of sliding displacement. Stability analyses of single spring and block models in which the frictional resistance of the sliding block obeys such constitutive laws show that sliding should always be stable if the frictional resis-tance increases with increasing sliding velocity (velocity strengthening), but can be either stable or unstable if the resistance decreases with increasing sliding velocity (velocity weakening). The nature of the behavior in the velocity weakening situation depends upon the ratio of the stiffness of the loading spring to a critical stiffness, which itself is a function of the constitutive parameters that describe the frictional surface. If this stiffness ratio is less than one then sliding will always be unstable. It the ratio is greater than one then whether sliding is stable or not depends on the mag-nitude of externally imposed perturbations of stress or loading velocity; sliding will be stable for small perturbations and unstable for large ones. One of the constitutive parameters that determines the value of the critical stiffness is a characteristic decay distance for the fading memory; thus in some senses this constitutive and stability behavior resembles slip weakening models. However, the constitutive law is richer than slip weakening models because it incorporates a description of how healing of the frictional surfaces occurs during the time when sliding is extremely slow between unstable slip events.

Predictions of the stability analyses agree with laboratory experimental observations as to
when sliding is stable and when it is unstable. Thus in a laboratory setting it is now possible to predict in some cases when unstable sliding will occur, although more data confirming the reliability of such predictions are needed. This enhances the confidence with which the laboratory constitutive laws and stability analysis can be used to understand the stability of sliding of faults and forms a mechanical basis for designing a program of earthquake prediction. Confidence in extrapolating the laboratory constitutive parameters to faults still depends upon knowledge of the way in which the characteristic decay distance in the constitutive law should be scaled, and so at present we can only make guesses as to the best value to use for this parameter to understand earthquakes. Work is currently underway to understand how to scale this parameter; at present the most likely possibility is that it depends on the uncorrelated roughness of the two sliding surfaces that in turn controls the size of the contact spots between the two surfaces. In the absence of knowledge of exactly how to do this scaling from the laboratory to faults, it is still possible to study the behavior of fault models by selecting a range of possible values of the characteristic decay distance. The recent model of this type presented by Tse and Rice (1986) exhibits behavior that is similar to that found or inferred for earthquakes: recurrence intervals are reasonable, locked patches exist in the seismogenic zone be-tween earthquakes, accelerating pre-seismic slip and deformation occur, initial dynamic rupture begins near the base of the seismogenic zone, dynamic stress drops and displacements are reasonable in magnitude, and post-seismic slip and deformation occur surrounding the dynamically ruptured patch.

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The hope that these studies hold for earthquake prediction on a variety of time scales is
that there are systematic changes in stress and displacements associated with all parts of the earthquake cycle and in particular there are accelerating changes in stress and displacement on the fault over a useful period of time prior to an earthquake. If sufficiently accurate measurements of these changes can be made, then it should be possible to predict earthquakes. The models can give us some insight into the spatial and temporal distribution of these premonitory signals and can provide an estimate of their magnitudes so that we can determine the feasibility of designing a field program to make the required measurements. Several valuable insights of importance to designing such a field progam are already available from such models. Computations of expected changes in strain and displacement during a model earthquake cycle show that: (1) It is not necessary to know the total stress in order to determine if an earthquake is about to occur since changes in stress are all that affect the stability criteria; (2) Points at the earth's surface undergo large changes in velocity within a year of a major earthquake and the pattern of the velocities can be used to infer accelerat-ing creep at depth, assuming the preparation zone is not too small; and (3) Continuous measure-ments of displacement and strain made in deep boreholes should be an important component of a field program since the largest premonitory changes occur nearest the eventual earthquake focus.

Introduction

Samples of rock that undergo episodic unstable frictional sliding in the laboratory exhibit accelerating slip prior to the unstable events if sufficiently precise measurements are made. Thus, although the term "stick-slip" descibes the fact that the surfaces alternately stick and slip, the term does not adequately describe the fact that slipping occurs over a very large range of velocities, and that even during the sticking part of the cycle very slow creep may occur. For years experimen-talists have hoped that the accelerating slip prior to an unstable event might provide a useful premonitory signal of the impending unstable event, and that if similar premonitory creep occurred on faults in the earth prior to earthquakes it might provide a basis for earthquake prediction. Obser-vations of such premonitory creep prior to earthquakes are not numerous, but some examples exist and lend support to the idea that such creep may provide a useful predictive tool.

Changes in stress and displacement for material around a fault occur at all times in the earthquake cycle and the changes in each of these vary considerably depending on which part of the cycle is considered. Hence stress and displacement are obvious first order quantities to monitor, as they are directly, rather than indirectly, related to the mechanics of the earthquake process. In order to design an effective field monitoring program it is necessary to have some reasonable es-timates as to the magnitudes of the changes to be expected in these quantities and their temporal and spatial distribution. The distribution and magnitude of these quantities are controlled by the interaction between the constitutive properties of the fault zone and the loading provided by the elastic distortion of the surrounding rocks. The stability of fault slip is controlled by the ratio of the stiffness of the elastically distorted surroundings to a critical stiffness which itself depends on the constitutive parameters of the fault zone.

The constitutive properties of the fault are still the least understood of the important
factors. The geometric irregularity of fault surfaces is an important factor; small scale irregularities might be considered as contributing to the constitutive properties whereas larger ones might be con-sidered explicitly in models that incorporate them into the elastic calculations. The role of the laboratory experiment is to elucidate the nature of the constitutive response of rock types that are found to be important constituents of fault zones. The constitutive data so obtained must then be incorporated into models of the faulting process that adequately represent the complexities of in-homogeneous slip under varying conditions on geometrically realistic faults. The laboratory experi-

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ments can play a further role, however, since the mechanical interaction that exists between the sliding rock surface and the elastically distorted loading apparatus is similar to that which exists on a much larger scale in the earth. Thus, the laboratory can serve as a testing ground for analyses of the stability of sliding. Most laboratory samples are sufficiently small that the two halves of the sliding rock sample move as nearly rigid blocks, whereas in the earth, dimensions are so much larger that the accumulation of elastic strain allows different amounts of slip along a single fault. Consequently, the analysis of the laboratory sample can be made using a simple one degree of freedom spring and slider-block model, whereas the elastic distortion of the rocks that load a fault distribute the spring contribution over the surrounding continuum. Nevertheless, the laboratory sample can be thought of as the most important fundamental element that exists in the natural setting, and if its behavior can be understood, then such elements can be linked together in theoretical analyses of fault behavior.

In this paper I briefly review results on the constitutive behavior of rock friction samples
in the laboratory, discuss the problems that must be solved in order to scale these results to natural faults, and, making some assumptions about how this scaling might be done, examine the patterns of stress and displacement adjacent to a strike-slip fault that are the expected consequences of using the laboratory data in a model of faulting. For the last topic I use the recent model of faulting of Tse and Rice (1986) that incorporates depth dependence of constitutive behavior. Taking their calcu-lated displacements on the fault as a function of time and depth, I further calculate the elastic stresses and displacements in the adjacent rock, focusing on surface measurements and depths that could be reached with monitoring wells.

Laboratory Findings

The coefficient of friction of those rocks that have been studied in detail varies with the
velocity of sliding and depends on the recent history of sliding through a memory that fades as a function of sliding displacement. Analysis of the stability of sliding for simple spring and block models using this constitutive behavior shows that it is this variation in frictional resistance with sliding velocity and displacement that determines the stability of sliding. The actual magnitude of stress or coefficient of friction play no role in controlling stability, although they do affect the mag-nitude of the energy dissipated by sliding on the fault. This simple observation shows that if the constitutive behavior of faults is similar to that of laboratory samples, then it is not necessary to measure the magnitude of the total in situ stress; measurement of changes in stress magnitudes and the associated changes in slip velocity are sufficient to determine in what part of an earthquake cycle the measurements are made.

Constitutive Behavior of a Frictional Rock Surface

Much of the frictional constitutive behavior of rocks in the laboratory can be understood
by study of the schematic diagram in Figure 1 and the associated equations. This idealization of actual behavior shows that upon an e-fold step increase in the velocity of sliding there is an im-mediate increase in the frictional resistance measured by a, and that with further sliding the resis-tance falls by an amount measured by b, where the magnitude of b may be less than or greater than a. Experiments show that for an e-fold step decrease in sliding velocity the response is simply the mirror image of that shown in the diagram; a sudden decrease in resistance of magnitude a which is followed by a gradual increase in resistance of magnitude b. If a-b is positive the response is termed velocity strengthening and if it is negative the response is velocity weakening. The decrease of mag-nitude b occurs over a characteristic distance of sliding, measured by L, which is the e-folding distance for the exponential decline. In some sense the response thus shows a displacement weaken-ing for an increase in sliding velocity. However, if the response is velocity strengthening, b less than

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a, then the net effect of the step increase in velocity is not a weakening with displacement. In the case of velocity weakening, it is useful to think of the resistance at lower sliding velocity as playing the role of classical static friction and the resitance at higher velocity as playing the role of classical dynamic friction. However, here the variation with velocity is continuous and in the idealization of the equations in Figure 1, effects of the same magnitude would be seen for a jump in velocity of the same ratio, regardless of the actual velocities. The constitutive behavior illustrated in Figure 1 is richer than displacement weakening since it also embodies a description of how strength is recovered as sliding slows down between possible high speed slip events.

The brief description above summarizes the experimental results of several workers in-
cluding Dieterich (1972, 1978, 1979, 1981), Ruina (1980), Dieterich and Conrad (1984), Okubo and Dieterich (1984), Weeks and Tullis (1985), Tullis and Weeks (1986a), and Lockner et al. (1986). Detailed support for the generalizations made in this section can be found in those references. Most of the experiments have been done on granite or quartzite, but some work on carbonates has also been done. Different mathematical representations of similar constitutive be-havior have been used by different workers. The one given in Figure 1 is a state variable representa-tion proposed by Ruina (1980, 1983) and is convenient because it has been used in nonlinear analyses of the stability of single degree of freedom systems by Gu et al. (1984), Rice and Gu (1983), Rice (1983) and Blanpied and Tullis (1986). Figure 1 represents the case in which only one state variable is used, but the formulation allows for any number. In practice, two state vari-ables are often required to accurately describe the behavior.

Recent experiments have been conducted over a wider range of sliding velocities than
used in the above studies, and the representation in Figure 1 in which the actual velocity does not play a role has been found to be inadequate. Dieterich (1978, 1979) proposed that at high velocity the dependence of the frictional resistance of granite on velocity disappears. More recent work on granite with much more complete data at high velocity shows that instead there is a switch from velocity weakening at low velocity to velocity strengthening at high velocity (Blanpied et al., 1986, 1987; Weeks, 1986, 1987), a type of behavior that has already been found for less brittle materials such as dolomite (Weeks and Tullis, 1984), halite (Shimamoto, 1986; Shimamoto and Logan, 1986), and calcite (Tullis and Weeks, 1986b). A transition back to velocity strengthening at even lower velocities has also been found for halite, suggesting that more common (and brittle) rock types may show this same effect at elevated temperatures. Most of the above quoted work on con-stitutive behavior has been done at room temperature, but some elevated temperature data has been obtained by Stesky (1978) and by Lockner et al. (1986). As more data over a wider range of conditions and greater variety of rock types become available and as their micro-mechanical processes become better understood, more general constitutive laws than the one presented in the equations of Figure 1 will be developed.

Little progress has been made in making microstructural observations that discriminate
between a varitey of micro-mechanical processes that may be responsible for the observed behavior. The relative importance of processes involving microfracturing, dislocation motion, pressure solu-tion, adhesion, stress corrosion cracking and hydrolytic weakening, for example, is not known. The physical significance of the state variables in the formulation of Figure 1 has not been identified, although there is some support for the suggestion of Dieterich (1979) that an important variable is the average lifetime of an asperity contact (Tullis, 1986). The importance of identifying the opera-tive micro-mechanical processes is that if the same processes can be shown to operate in both the laboratory and the earth, then the validity of extrapolating the laboratory results to earthquakes is enhanced.

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In summary, over a limited range of sliding velocities (0.001 to 10 u.m/sec), several important rock types at room temperature show velocity weakening behavior (Figure 1, with b greater than a), but that at higher and lower velocities the behavior may change to velocity stren-ghening. Velocity strengthening is predominant at sufficiently high temperature. This statement is based on only a limited quantity of data on only a few rock types and so applying it to the earth should be done with caution until more data become available. An important qualification is that some experimentalists (Solberg and Byerlee, 1984; Byerlee and Vaughn, 1984; Lockner et al., 1986) find velocity strengthening under conditions where others find velocity weakening. The ex-planation for this discrepancy is not yet known.

Stability of Sliding

Our understanding of the stability of sliding in the laboratory has progressesd con-
siderably in the past few years, following the demonstration by Dieterich (1978) of a quantitative relation between the stiffness of the loading system and the magnitude of parameters in the friction constititive description. We now have a good understanding of the transition between unstable slid-ing (stick-slip) and stable sliding in the laboratory. This is best demonstrated by the results of Tullis and Weeks (1986a) and of Blanpied and Tullis (1986) that will be described briefly below, since they offer some insight into how laboratory data might be used to understand the stability of sliding in the earth and to predict earthquakes.

The stability analysis used is that of Gu et al. (1984), Rice and Gu (1983) and Blanpied
and Tullis (1986); it deals with systems with one degree of freedom, i.e. single block and spring models. This analysis cannot be applied directly to the more complex coupling of natural faults to the source of their loading, but it applies very well to laboratory samples that are small enough that they slide as rigid blocks. The expected behavior of such systems for a material that shows velocity weakening is summarized in the phase plane plot of Figure 2, taken from Rice and Gu (1983). The actual behavior of an experimental sample (Figure 3) shows that in the only case tested in detail to date, the stability analysis accurately describes the behavior of the experimental system (Tullis and Weeks, 1986a). The fact that a three-fold externally imposed perturbation in the loading velocity would cause an instability was predictable from knowledge of the constitutive parameters. An even more detailed analysis of this situation has been given by Blanpied and Tullis (1986) who take into account the role of the second state variable that is needed to accurately describe the behavior.

There are three possible cases that can exist concerning stability. In the first case, which
occurs for a velocity strengthening material, sliding should always be stable and perturbations im-posed on a system will decay. In the second case, illustrated in Figures 2 and 3, sliding can be either stable or unstable, depending upon the size of the externally imposed perturbation. In the third case, sliding can only be unstable. In the single degree of freedom situation, the factor determining which case applies is the ratio of a system stiffness K to a critical stiffness Kc. The conditions are most easily described for a one state variable system, but the general behavior is similar for more than one state variable. Case three, only unstable sliding possible, corresponds to K/Kc < 1, and the behavior of the stress, velocity, and displacement through a cycle is shown in Figure 4 taken from Rice and Tse (1986). Case two, conditional stability, corresponds to K/Kc > 1. Case one, always stable sliding, corresponds to Kc < 1.

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Application to Natural Faults

The analysis of stability and its success in making a "prediction" (so far only retrospec-
tive) of unstable sliding in the laboratory offers several possibilities for application to earthquakes on natural faults. If the constitutive parameters for a fault can be determined and appropriate account taken of the inhomogenoeus elastic deformation of the earth that is loading that fault, then it should be possible to determine to which of the three above cases the fault corresponds. Experience sug-gests that the central creeping portion of the San Andreas falls into case one (always stable) and that the "locked" portions north and south of there fall into case three (always unstable). Experience from laboratory experiments suggests that the differences may result from the former region being dominated by velocity strengthening material and the latter regions being dominated by velocity weakening material. This expresses mechanically the suggestion made by Alien (1968) that the creeping part may be dominated by serpentine in the Franciscan formation and the "locked" parts dominated by granitic basement. Data on the velocity dependence of serpentine are not yet avail-able, but other aspects of its characteristics suggest it may be velocity strengthening. Geographically intermediate regions may fall into case two, so that small perturbations may not grow into earthquakes, but larger ones may.

Given this background, the task of earthquake prediction involves more than just know-
ing to which case a given region corresponds. Knowing the timing and the size of an earthquake requires more detailed knowledge of the temporal and spatial variation of stress and velocities. For simple independantly acting faults falling into case three, perhaps such as the Parkfield segment of the San Andreas fault, monitoring of the stress and velocity field through an earthquake cycle should allow determination of what magnitudes of changes are associated with each distinctive part of the cycle shown in Figure 4. If such details were known, then it should be possible to predict the effects that a remote earthquake, such as the Coalinga earthquake of May 2, 1983, would have on modifying the normal timing of a characteristic earthquake. For large earthquakes on the northern and southern San Andreas fault, knowledge of where today's date falls in the cycle could be greatly aided by combined measurements of stress changes and velocity changes through time. If the laboratory constitutive parameters can be applied to these faults, then we can predict what mag-nitude of changes we should expect in different parts of the cycle.

How to Scale the Constitutive Parameters

To the layman it may seem impossible that measurements made on small laboratory
samples can have much relevance to the mechanics of sliding on crustal scale faults, and to the earth scientist this also sometimes seems impossible. Certainly it cannot be taken for granted that the measurements are relevant, but it should be a testable proposition. If the processes that are found to occur on the spatial and temporal scales of the earth also occur on the laboratory scales, then careful work should allow the appropriate extrapolations to be done. The situation is similar to that in the field of high temperature ductile rock deformation, which is ahead of the field of rock friction in this regard. In ductile rock deformation, many of the important processes operating in both the field and the laboratory have been identified and their dependencies on temperature, dif-ferential stress, pressure, grain size, and chemical environment are known, at least in part, and are being investigated further. Those natural processes that can be studied readily in the laboratory, such as dislocation creep, are reasonably well understood. Those natural processes that operate too slowly for convenient laboratory study, such as pressure solution, are much less well understood. At present, since we do not know much concerning the details of the micro-processes that operate in frictional sliding in either the earth or the laboratory, it is not easy to convincingly argue for the direct relevance of the laboratory measurements.

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However, at least dimensional arguments can be used to support the idea that the dimensionless constitutive parameters a and b determined in the laboratory may not require scaling for application to the earth. If the values of these parameters depend on time and/or physical and chemical conditions, then direct application to the earth requires knowledge of these dependencies. The situation with the constitutive paremeter L is not as simple, however, since it has dimensions of length, and the length scales of the laboratory and of the earth are vastly different. What is needed is to make the appropriate non-dimensionalization of L, but this requires knowledge of what physi-cal length controls the magnitude of L.

The size of L is most likely controlled by the diameter of spots that are in contact across
a frictional sliding surface, as has been suggested by Dieterich (1979). This is similar to the concept of adhesional bonds in the friction of metals, and although the behavior of brittle silicate minerals at room temperature may be different in detail, the general ideas may still apply. Dieterich (1979) and Okubo and Dieterich (1984) found that there was a general correlation between the roughness of ground surfaces and the magnitude of characteristic decay distances in the frictional constitutive law. The idea has additional support from recent calculations of the diameter of contact spots be-tween two surfaces of measured roughness. For four different roughnesses the inferred diameters of contact spots differ from one another and are of approximately the same size as the characteristic decay distances measured for these roughnesses (Tullis, 1986). These data suggest that measure-ment of the roughness of natural fault surfaces may provide a way to calculate the diameter of the spots where opposing fault surfaces touch and thus provide a basis for scaling the parameter L from the laboratory to the earth. Measurements of fault roughness undertaken for this purpose have been made (Power et al., 1986) and more are underway.

In the absence of actual knowledge of how to scale the laboratory-based constitutive parameters to the earth, it is possible to try several values for the parameters and see if reasonable behavior is predicted. This approach has been taken by Tse and Rice (1986) who use the laboratory values of a and b at room temperature and in the temperature range of 300-700 °C to construct a likely set of values for a and b from the earth's surface to 30 km depth. Lacking knowledge of the best values for L, they try several values and study the way the resulting behavior depends on the value chosen.

Fault Model

The model of Tse and Rice (1986) of the mechanical behavior of a strike slip fault is
the best one that currently exists in the sense that it uses laboratory constitutive information in a detailed numerical simulation of expected behavior. In their model the constitutive parameters and the slip vary as a function of depth, but have no variation along strike. The earthquakes occur spontaneously in this model and are quite realistic in terms of the magnitudes of the stress drops, co-seismic displacements, depths of rupture initiation, pre- and post-seismic creep, etc. Figure 5 shows an example of the predicted displacements at the fault surface from this model and Figure 6 shows the distribution of slip velocity on the fault at various times. Although the behavior of their model fault may not exactly replicate the behavior of real faults in actual earthquakes, it appears to come quite close, and numerical models can be refined in the future to take into account both better constitutive information as it becomes available and more complex fault geometries.

The model of Tse and Rice (1986), and more complex ones that may succeed it, are
necessary to take into account the variation in constitutive properties that can occur as a function of

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both depth and rock type along strike, and the variation that can occur in the amount of slip due to stress inhomogeneities. This situation is geometrically more complex than a single spring and block and it is not possible to describe the behavior of the entire system by a single trajectory on a phase plane plot of the types shown in Figures 2-4. Nevertheless, there will be calculable temporal changes in both stress and velocity at all points influenced by the earthquake cycle in the model, and these may resemble the actual variations that occur in the earth adjacent to active faults.

Expected Changes of Measurable Parameters During the Earthquake Cycle

In order to design a rational program of earthquake prediction, it is important to know
the best places to make measurements of those quantities whose changes can signal a coming earthquake. In addition it is important to know how large the signals can be expected to be, in order to determine whether they can be detected with existing or future instruments. The principal problem of earthquake prediction is the difficulty of placing a dense enough array of the appropriate instruments at focal depths. If we had complete knowledge of the stress and displacement field we could probably predict nearly all earthquakes, but with only the incomplete knowledge available it is a much more difficult task. A good model of what to expect can provide much insight into how to deploy the appropriate instruments. In the remainder of this paper I will look at what changes in various physically measurable quantities are predicted by the model of Tse and Rice (1986). Similar and more detailed calculations could be done in the future if improved models become available.

Method. I have used data of the type illustrated in Figures 5 and 6 as input to calculate
the distribution of strain, displacement, strain rate, and velocity away from the fault, both at the earth's surface and at depth. This was done using the boundary element displacement discontinuity method (Crouch and Starfield, 1983), taking the distributions first of displacement and then of velocity on the fault as input. The displacement and velocity distributions were each treated as a series of screw dislocations. The 30 km thick lithosphere was divided into 128 elements and the values of the displacement or velocity dislocation for each of these elements were taken directly from the computer tape that was output from the CRAY I computer in the study of Tse and Rice (1986). The Green's functions for the two non-zero strain and strain rate components and for the displacement and velocity at each location away from the fault were taken from equations A3b and A5 of Tse and Rice (1986). Since equation A3b contains a rigid displacement this was eliminated by replacing the numerator of the fraction by the expression {exp[ip((z-a) + iy)/H] - 1}, where the terms have the same meaning as in Tse and Rice (1986). Following Tse (1986), a term that takes into account the uniform straining of the region was added to the contribution from the distribution of dislocations. The contribution from the uniform straining is given by the first terms of equations 5.2a and 5.3a of Tse (1986), except that the plate velocity Vpl is replaced by the difference be-tween Vpl and the thickness average of the dislocations over the fault surface, a minor error in those equations. All of the boundary element computations and plotting of results were done using an IBM RT-PC.

Results. Changes in strain and displacement prior to the earthquake are most obvious at
the focal depth because it is there that the fault is most unstable. To some extent the fault both at depth and at the surface behaves like separate single-degree-of-freedom slider blocks that have dif-ferent ratios of stiffness to critical stiffness and thus have different individual stability tendencies. The individual tendencies of the two different regions progressively diverge as an earthquake ap-proaches and because of this there are increasing stresses resulting from the elastic coupling between the different depths that cause the surface to be progressively more affected by the changes occur-ring at depth. Hence although the surface would not become unstable at the time of the earthquake were it not for the instability occurring at depth, changes do occur at the surface in response to the

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growing instability at depth. As is discussed below, the changes in stress that occur at the surface are of insufficient magnitude and occur over the wrong time scale to be very useful for earthquake prediction, but the rate of changes in displacement that occur at the surface should allow a predic-tion to be made. However, a much better prediction could be made if measurements of strain and displacement can be made at depth in wells drilled about 5 km deep. This not only would allow a greater confidence level to be placed on the prediction, it also would allow an earlier warning of the impending earthquake.

The exact depth at which the measurements are best made depend on the depth at
which the fault is most unstable. In the model of Tse and Rice (1986) shown in Figures 5 and 6 the accelerating premonitory slip and accompanying strain reduction is most pronounced at a depth of 5 km until a few minutes before the earthquake when the most rapid motion shifts to a depth of 6.5 km which would be the hypocentral depth of the earthquake. The particular depth at which these events occur depends on the variation of the constitutive parameters used in the model and so the exact depth variation of behavior reported in the remainder of this paper would change if different depth variation of the parameters was used. Although they may be somewhat shallow, the depths used herein are not unreasonable for strike slip earthquakes on the San Andreas fault in California. This is illustrated by the facts that the focus for the 1966 Parkfield earthquake mainshock was about 9 km deep (Poley et al., 1987), that the maximum slip was on a patch centered at about 5 km deep for this earthquake (Stuart et al., 1985), and that the depth distribution of coseismic slip for the 1906 San Francisco earthquake (Thatcher, 1975) was similar to the coseismic slip shown in Figure 5 for the model earthquake.

The most useful overview of the nature of the earthquake cycle at a variety of depths is
provided by Figure 7, which is a phase plane plot for five different depths at a distance of 500 m away from the fault plane. In this plot the shear strain parallel to the fault plane in the slip direction is plotted versus the natural logarithm of the velocity of points moving in the slip direction relative to the center of the fault (the half rate), normalized by the remote velocity of one plate relative to the other (the whole rate, taken as 35 mm per year). The shear strain can be converted to shear stress by multplying by the shear modulus of 3 x 104 MPa used by Tse and Rice (1986). Selected times before the next earthquake are labeled on Figure 1.

Careful study of Figure 7 reveals many interesting facts concerning the earthquake cycle.
The stress changes during a cycle are progressively larger at depth due to the simple fact that the increasing normal stress causes larger changes in frictional slip resistance. Less obviously on initial inspection, the temporal increase in the shear stress following the previous earthquake ceases much earlier at the surface than it does at depth. This is more easily seen in Figure 8 where it is clear that the strain at the surface near the fault zone ceases to increase in the last 20 years prior to the next earthquake, whereas the strain at depth continues to build until less than a year prior to the earthquake. Also during the final year prior to the earthquake the changes in strain at the surface are quite small compared to those at depth. This can be seen from Figure 7 by comparing the changes that occur between 0.8 years and 1 minute prior to the earthquake at different depths. This is also well shown in Figure 9 which shows the difference between the shear strain as a function of depth for 0.8 years and 18 hours prior to the earthquake. This figure clearly illustrates how the failure process is restricted to depths below about 3 km even only 18 hours prior to the event. Because the failure begins very close to 5 km deep, the phase plane plot for that depth in Figure 7 shows a larger decrease in shear stress prior to the dynamic rupture than do the plots for depths of 4 and 6 km.

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Although the shear strain changes at the surface are not very useful for prediction of the earthquake, Figure 7 shows that the velocities at the surface as well as at depth will increase rapidly prior to the earthquake and could be useful for prediction purposes. It is also apparent, however, that the velocities at the surface lag those at depth and so less warning is provided by having only surface measurements.

Discussion. In the following discussion I will first focus on the changes that occur in
shear strain and how these compare with the magnitude of strain changes that can be made with existing instruments. Then I will turn to a similar discussion of the changes in displacements and velocities.

Shear strain changes predicted at depth by this model are large enough that they should
be recognizable from background noise and other tectonic signals, but those at the surface at any distance from the fault are too small to be recognized. The long term change in strain is shown in Figure 8 that already has been discussed. The fact that the strain continues to increase at depth to within a year of the earthquake means that the cessation of this rise is a very useful intermediate term earthquake predictor. At the surface the strain did not continue to increase for the last part of the earthquake cycle because some slip is occurring at the surface that prevents the stress from building up. This is shown in Figure 10 where the contrast between the more rapidly slipping surface and the very slowly slipping deep region is illustrated.

Whether a given strain change is detectable depends both on the sensitivity of measuring instruments and the ambient noise level. Figure 11 shows the calculated changes in strain for two surface locations and one at depth, superimposed on a plot from Sacks and Linde (1987) of changes in strain measured near the San Andreas fault by two different types of borehole strainmeters. The strain measurements are residuals, after drift due to the recent installation of the instruments has been removed. One measurement is made by a strainmeter that only measures volumetric strain while the other is from an instrument that measures shear strains as well (Gladwin, 1986; Gladwin et al., 1987). These instruments are located several km apart and both show a tectonic signal in response to the April 24, 1984 Morgan Hill earthquake which occurred about 50 km away at the time on the record marked MH. The strain event associated with this earthquake persisted for 6 to 9 months at both stations. The background level on the shear instrument, upon which this tectonic signal is located, gives some idea of the level of background noise that may be expected from such installations. Some improvement may be expected with time as the instruments "settle in" and methods are devised for filtering out some of the noise, but the detection levels of the instruments in field installations will never approach their intrinsic sensitivities.

The size of the strain changes calculated for our model earthquake are compared to measurable strain changes in Figure 11. For the model calculations the earthquake is assumed to occur at the time corresponding to the right side of the figure. The changes at the ground surface at both distances from the model fault are too small to detect until the last 1 to 3 months. On the other hand the changes at depth are easily detectable over the entire time period and become extremely large during the last three months. Obviously separating the tectonic signal due to the impending earthquake from other tectonic signals such as the remote Morgan Hill earthquake is necessary. This would have to be done by having an adequate distribution of stations to determine the spatial extent of the signal and should be done with knowledge of expected coseismic changes from known remote earthquakes. A useful strategy would be to deploy instruments both at the surface and at depth since they might both record similar signals from remote events and their signals could be differenced to study the nature of local effects; this procedure might also remove some types of noise.

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Changes in displacement and velocity derived from them are large both at the surface
and at depth prior to the model earthquake. The breaking of an irrigation pipeline spanning the San Andreas fault trace about 9 hours before the 1966 Parkfield earthquake (Bakun and Lindh, 1985) suggests that accelerating fault slip does sometimes occur prior to real earthquakes. The fact that this location is at the surface and about 18 km southeast of the epicenter of the main shock which had a hypocentral depth of 9 km suggests that, just as for the model earthquake, accelerated creep may not be restricted to the focal area.

An important result of the model is that as the earthquake approaches, changes in the distribution of velocities at the surface occur as a function of distance from the fault. These various changes should allow prediction of an impending earthquake, since the magnitudes of the calculated changes are such that they can easily be measured with existing creep meters and laser geodimeters. As the earthquake approaches, the strain within several km of the fault becomes increasingly non-uniform. Consequently the fact that an earthquake is approaching can be more clearly recognized if homogeneous strain is not assumed when reducing line length data for lines reaching different dis-tances from the fault; it is most useful to covert this line length data into velocities as a function of distance from the fault.

Measured displacements will be used to determine velocities which themselves are more
useful for understanding in what part of the earthquake cycle observations are being made. Distance measurements can be made to about one part in 107 of a line length with multi-wavelength laser instruments (Langbein, 1986) so that for lines a few km long displacements can be measured to a few tenths of a millimeter; creepmeters across fault traces can measure to about an order of mag-nitude better than that. Assuming that measurements can be made to within 0.2 mm for many lines of intersest, it would take about (0.2/R) years to detect a change in rate of R mm/yr. Thus, it should be possible to detect changes in velocity of nearly 1 mm per year in a time interval of 2-3 months, changes of 10 mm per year in a time interval of about one week, and changes of 100 mm per year in about 1 day.

The predicted changes in velocity from the model fault are large enough during the two
years prior to the earthquake that they would be detectable given the above resolution, and during the last few months the changes become extremely large. These predicted changes in velocity are shown in various ways in Figures 12, 13, 14 and 15. The velocities are plotted versus time until the earthquake in Figure 12. Study of this figure shows that these changes should be easily detectable. As the earthquake approaches it is more useful to plot the time to the earthquake on a log scale as has been done in Figures 13 and 14. In Figure 13 the velocity itself is plotted, and in Figure 14 the log of the velocity; for points 3 km from the fault, velocity changes of about 200 mm per year occur a week before the earthquake but are considerably smaller closer to the fault as comparison of the curves for 1 m, 500 m, and 3000 m show. This is a reflection of the fact that points at the ground surface some distance from the fault are more sensitive to the displacements occurring at depths of about 5 km on the fault than are surface points near the fault. At the same time that the velocities are increasing everywhere, the peak in velocity at the surface is shifting from the trace of the fault to points progressively farther from the fault, as is clearly shown in Figure 15 for the interval from 0.8 years to 0.8 months. At 18 hours before the earthquake the highest surface velocities are at 4.5 km from the fault and are about 560 mm per year. This change in the pattern of peak velocity is a manifestation of the fact that the most rapid increases in velocity are occurring at depth due to the growing instability there and this change in the pattern is itself a useful confirmation of the impend-ing earthquake.

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Although the increase in the velocities at the surface and the change in their pattern
should allow the earthquake to be predicted without measurements of velocity made at depth, the velocities at depth begin their dramatic increase earlier than do those at the surface (Figures 12 and
13) and attain much higher values (Figure 14). Consequently in order to obtain an earlier warning and to provide some measurements that are complementary to those at the surface, it would be very useful in a program of earthquake prediction to be able to monitor the velocity of points within deep wells as a function of time and depth. A measurement technique with a geometry resembling an allignment array, but running down a drill hole, would be extremely valuable, especially if it could be monitored automatically.

Aki (1986) has suggested that the ideal earthquake prediction is one which gives the probability rate of an earthquake to be 1/t at time t prior to the earthquake. Thus, 1 year prior to the earthquake 1 earthquake per year is predicted, 1 hour prior to the earthquake 1 per hour is predicted, etc. In this case the log of the probability of an earthquake's occurrence in a given interval of lime increases linearly when plotted against minus the log of the time interval to the earthquake, and the value of the probability rate is 1 at that given time interval prior to the earthquake. Thus, as the time of the earthquake approaches, the strengths of the signals used to predict it grow very rapidly and the certainty of the earthquake's occurrence goes up with the size of the signals. Large slip velocities in the earth, such as are predicted in the deterministic fault model of Tse and Rice (1986), have high probabilities of being associated with an earthquake, and as shown in Figure 14 the log of the velocity is nearly linear when plotted versus minus the log of time to the next earthquake. If a more rigorous association could be made for the earth between slip velocity and the probability of an earthquake occurring, then this premonitory slip velocity might come close to fulfilling Aki's requirement for an ideal earthquake prediction tool. The plot for 5 km depth appears to provide a better prediction since it is closer to being linear over the range of times from several years prior to the earthquake to 1 hour before the earthquake, and because higher absolute velocities should be associated with higher probabilities of an earthquake. The sur-face data may be equally good except that they show slope changes much closer in time to the earthquake; the surface data at a distance 3 km from the fault seem better by this criterion than those closer to the fault.

Whether or not slip velocities can be converted into probabilities of an earthquake's occurrence, the rapid increase in velocity illustrated in Figures 12-15 would be an extremely useful earthquake prediction tool if it occurs commonly in real earthquakes. When velocities become very high, the only other possibility is that a creep event is underway that will stabilize without becoming an earthquake. Prior history of the nature of slip of a given fault segment is perhaps the best way to determine the likelihood of the event's being an earthquake versus a creep event. Tullis and Weeks (1986) suggested that the upward concavity of a phase plane plot such as the one of Figure 7 for a depth of 5 km is diagnostic of a runaway instability; unfortunately the time scale over which that occurs is too short to be very useful as a prediction, even if instruments were capable of detecting it.

Conclusions

Understanding of the constitutive behavior of rocks in the laboratory has advanced to
the point that there is some value in exploring the behavior that natural faults would show if they followed the same constitutive behavior. Although a large number of uncertainties exist concerning the extrapolation of the laboratory data to natural faults, fault models such as those of Tse and Rice (1986) show behavior that matches the behavior of natural faults undergoing earthquake cycles in a remarkable number of details. Making the assumption that a real fault will behave in exactly this

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way, the model can be used to determine the distribution of strain and velocity that should exist during an earthquake cycle and these calculations can be used to guide the location and character of field instruments that will be used to attempt prediction of earthquakes.

Clearly the suggestions from this modeling are subject to a number of limitations. The
fault model on which they are based has no variation along strike and hence the size of the signals that exist at the surface are larger than would be the case if the focal region consisted of only a patch rather than a long strip. Furthermore the calculated strains and velocities will differ if other con-stitutive parameters are used. Without doing additional calculations with other combinations of paramerters is it not easy to say how strongly the calculations depend on the particular choice of parameters chosen. In general the qualitative behavior will be similar for a range of likely parameter values, but the magnitue of the effects and the time scale prior to the earthquake over which they occur will differ. The fact that this set of parameters predicts some creep at the surface clearly shows that the model is not an accurate description of the locked parts of the San Andreas fault. Model parameters probably could be chosen such that the surface creep would be small enough that it could be overlooked in a field monitoring program. Unless ad hoc assumptions were made about the depth variations of the constitutive parameters, such a choice of parameters would result in an earthquake that would have smaller precursory velocities at all depths and the time interval over which changes in velocity and strain occur prior to the instability would be smaller. They could be so small as to be of no use in a practical prediction program. On the other hand, the lack of creep at the surface means that more stress would build up at the surface than in the specific model studied here, and so monitoring of strain at the surface could be of more value than suggested by the current calculations. Exploration of the actual consequenses of changing the model parameters must await further calculations.

Given the above cautionary comments, it is nevertheless quite useful to consider the implications of the model calculations. Not surprisingly, they indicate that if it were possible to deploy instruments at focal depths in a series of wells, the sensitivity to the premonitory changes would be enormously increased. This will be even more true for the more realistic situation in which slip on the fault initiates on a patch rather than an along-strike strip. Strain changes at the surface do not appear to be large enough to be useful in intermediate term earthquake prediction unless the focus is much shallower than in this model, but if they could be employed at focal depths they would be extremely useful. Precise methods of surveying the bottom of wells should also be devised, since the displacements at depth are also very useful in prediction. Measurements of velocities at the surface can be quite sensitive to a coming earthquake even if the focus is 5 km deep. The spatial distribution of the velocities should be carefully recorded and plotted; if homogeneous strain is assumed within 5 km of the fault zone, significant details of the displacement pattern will be missed that can be very useful in detecting precursory slip at depth.

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746

V,

b-a

= eV1

aln (v;) + b
a-b

displacement

Figure 1. The response of friction to an abrupt increase in sliding velocity of magnitude
e (- 2.7183), applied while the system is sliding at steady state with coefficient of friction ji,. The coefficient of friction |JL undergoes a transient change in velocity of magnitude a in the same sense as the velocity change. For a constitutive law with one state variable \\r, with continued sliding at velocity V2, the value of bxj;, and consequently JJL, decays by an amount b (for the e-fold increase in velocity) to a new steady state level (jL2. The decay occurs exponentially as a function of slip displacement, falling lo an amount bA\J//e above the new steady state level after a displacement L.

747

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w

0)

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0)

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2

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0

- 1

-2

-3

-4

- 5

»

Unstable region

\

I Stable region \

/ ^v

V

I I
-4 -3 -2 -1 0 6

In (dimensionless slip speed), Iny

Figure 2. Phase plane plot showing stability boundary, steady state line, and slip be-
havior trajectories for samples starting at two different states (from Rice and Gu, 1983). Horizon-tal axis is ln(slip velocity) normalized to an arbitrary velocity V*, in this case chosen to be equal to the load point velocity V0. Vertical axis is dimensionless shear stress, in this case coefficient of friction, normalized by the magnitude of the direct velocity effect. This plot is drawn for a par-ticular system stiffness and combination of the one state variable constitutive parameters a, b, and L. If the starting point in terms of surface slip velocity and friction is below the stability boundary, the system response is to spiral toward a new steady state at the point on the steady state line where V = V0. If the starting point is above the boundary, the result is unstable sliding as shown by the trajectory going to infinite velocity (neglecting inertia). The spirals made by the stable trajectory indicate an oscillatory approach to steady state on a friction vs. displacement plot. A load point velocity jump beginning at steady state starts at low velocity and evolves toward a higher velocity; consequently, on this plot it corresponds to a trajectory starting to the left and above the origin which is the position of the load point velocity after the velocity jump. Conversely, a load point velocity decrease beginning at steady state starts to the right and below the origin.

748

Predicted Stability Boundary for b-a =.00476, a = .006, L = 7 /um

Observed Experimental Behavior

.64-,
V*/V = 2

V*/V = 3

i r
2 0 246
In (V/V*)

50 fj,m Displacement

Figure 3. Comparison between predicted and observed stability on plots of the type
shown in Figure 2. On the left are plots of data, together with calculated stability boundaries. On the right are the same data plotted versus displacement. On the top, for a velocity jump of 2 to 1, the theory predicts stable sliding while for a jump of 3 to 1 (below) unstable behavior is predicted. The "X" marks the predicted steady state point; the data fall below it because of a long term trend that a constitutive law with only one state variable does not account for (see Tullis and Weeks, 1986a). The machine stiffness k used to calculate the stability boundary is 0.0708 MPa/|j-m, which yields a normalized stiffness K (equal to kL/a) of 2.590, and a ratio of K/Kc - 1.53. The data point marked by the question mark could lie much farther to the right because the sampling interval of one second cannot resolve high velocities.

749

30-*-

Depth (km)

AtoB
BtoC
CtoC
CtoD
DtoE
EtoF
FtoG
GtoII

Time Interval

2.4 yrs.
4.6 hrs.
instantaneous
4 sees.
3.5 mins.
77.8 days
3.6 yrs.
77.2 yrs.

Tcy = 83.8 yrs.

C2L40LOO

Figure 5. Displacement versus depth for three earthquake cycles from model C2L40LOO of Tse and Rice (1986, Figure 7), in which the time interval between earthquakes is 83.8 years. Each line represents a constant time, but the time interval between the lines used for illustration varies during the earthquake cycle because much of the interesting behavior occurs just before or just after each earthquake. The time intervals between some representative lines are given. The earthquake occurs between lines C and C'. Surface creep occurs during the entire earthquake cycle, but the accelerating premonitory slip is greatest at about 5 km depth, as a result of the increasing normal stress with depth and the temperature dependence of the constitutive parameters a and b. The value used for L in this model is 40 mm.

751

C2L40LOO

Time Interval

AtoB
BtoC
CtoD
DtoE
EtoF
FtoG
GtoH
Htol
ItoJ
JtoK

2.5 sees.

1.2 days
76.6 days
3.6 yrs.
24.1 yrs.
42.4 yrs.
11.7 yrs.
1.7 yrs.
1.3 hrs.
4.2 sees.

Tcy = 83.8 yrs.

Depth (km)

Figure 6. Log normalized slip velocity versus depth during one of the earthquake cycles shown in Figure 5 (Tse and Rice, 1986, Figure 8). The lines correspond to constant times, with intervals between them as indicated; the times, time intervals, and labeling are different from those in Figure 5. The focal depth is 6.5 km as shown by time K just prior to the earthquake (time A), during which the slip velocity is set to a limiting value. Vpl is the remote relative plate velocity, taken as 35 mm per year.

752

Shear strain

0 ««tt.
+ dirt.
£ (Hit.
a di.t.
X dirt.

900 m; d»ptti » 6000 m
500 mt d»ptti * 5000 m
500 m; dfpfh « 4000 m
500 mt d»ptti 2000 m
500 m; d»pth = Om

J-H
+->

</>

8
o

- 4-20 2 4 6 8 10 12 14 16 18 20

In (V/Vpl)

Figure 7. Phase plane plot for points at a variety of depths, 500 m from the fault, for.
an entire earthquake cycle, computed from the displacements and velocities in the model of Tse and Rice (1986) that are illustrated in Figures 5 and 6. V represents the velocity of either side of the fault with respect to the symmetry point at the fault so here it is half of the relative velocity of points 1 km apart, 500 m on either side of the fault; Vpl represents the total relative plate velocity which is taken to be 35 mm per year. For the boundary element calculations only 45 repre-sentative times have been selected from the much larger number needed by Tse and Rice to model the instability. Times prior to the earthquake for the last part of this sequence of synchronous data points are 28.4, 20, 12.2, 8.1, and 2.8 yr, 297, 174, 88, and 24 days, 17.5 hr, 32, 3.3, and 1 min, and 26, 14, 8.4, 5.6, 3.8, 2.6, 2.0, and 1.4 sec.

753

Shear strain

140

120

100

.5 80
03

60
IN

40

20

0

-20

0 dttt. 500 mj d«pth - 5000 m
+ dt*t. 500 m; d«pth a Om
X dtf t. « 5000 mi d«plh = 0 m

90 80 70 60 50 40 30 20 10 0

Time to next earthquake (years)

Figure 8. Variation of shear strain as a function of time over the entire earthquake cycle at three different locations. Details of these curves for the 1.4 years prior to the earthquake are shown on an expanded scale in Figure 11.

754

Shear strain

140

120

100

2 80
o

60 h-

40

20
0

0 dttf. = 500 m; Mm* = -207.24 day*
+ ditt. - 500 m; llmt - -17.5J hour*

2000 4000

Depth (meters)

6000 8000 10000

Figure 9. Comparison of the magnitude of the shear strain at two times prior to the earthquake shows the depth interval over which the initial instability develops.

755

Displacement (meters)

0.55

0.50

0.45

0.40

0.35

0.30

0 dirt. « 1 m; d»pfh = 5000 m
+ dtil. - 1 m; d»ptti « Dm

90 80 70 60 50 40 30 20 10 0

Time to next earthquake (years)

Figure 10. Much more displacement on the fault occurs between earthquakes at the
surface than at the focal depth. Zero displacement is defined just after the dynamic slip in the previous event. The displacement after the impending earthquake at all locations is 1.468 m. This is displacement of either side of the fault with respect to the symmetry point at the fault; the net slip on the fault is twice this displacement.

756

0

cd

o
!

O

- 2

residuals: shear at SJB, dilatation at SRL

Dist. = 500 m;
Depth = 5000 m
Dist. = 5000 m; v
Depth = 0 m \

Dist. = 500 m; _X
Depth = 0 m

Figure 11. Shear strain changes measured by shallow borehole strainmeters near the
San Andreas fault (irregular solid lines, see text for discussion) compared with predicted mag-nitudes of premonitory strain changes for the model earthquake cycle (dashed and dotted lines). The time of the model earthquake is the right edge of the plot. The three theoretical curves are the same as those of Figure 8 at an expanded scale and with arbitrary zero shifts along the strain axis to show them all on the same plot.

757

Velocity (meters per year)

0.03500 r-

0.03000

0.02500

0.02000

0.01500

0.01000

0.00500

0.00000

0 dirt. = 500 mi dvpth * 5000 m
+ di«l. « 5000 m; dvplh = Om
X dirt. = 500 mi dvpth = Dm
LJ did. « 1 m; depth B 0 m

4.00 3.50 3.00 2.50 2.00 1.50 1.00 0.50

Time to next earthquake (years)

0.00

Figure 12. Illustration that the velocity at the focal depth is lower than that at the
surface until about 6 months prior to the earthquake and then becomes larger than that at the surface. More detail during the 6 months prior to the earthquake is shown in Figure 13 and 14. Velocities are with respect to symmetry point at the fault as defined in caption to Figure 7.

758

Velocity (meters per year)

1.00 r-

0.90

0.80

0 did. » 500 m: depth » 5000
T dttt. = 9000 m: depth = Om
)K dlft. « 500 m; depth » 0 m
D dl«t. = I m; depth = 0 m

10 Yr. 1 Yr. 1 Mo. 1 Wk. 1 Day

A
1 Hr.

A
1 Min.

1 0-1-2 -3 -4 -5

Log Time To Next Earthquake

- 6

Figure 13. Velocity at different locations using a log scale for time to emphasize the
period of time prior to the earthquake. The increase at depth is the largest and occurs the earliest. Within a few days of the earthquake the increase at the surface at a distance of 3 km is more dramatic than closer to the fault. This reflects an accentuation of the pattern of velocity that begins about 6 months prior to the earthquake as shown in Figure 15. Velocities are with respect to symmetry point at the fault as discussed in caption to Figure 7.

759

10

8

c^ 6

ex I'
£ 2

0

- 2

-4

0 dl»t. * 500 m; depth = 5000 m
+ dltt. « 3000 n»t depth = Om
X dltt. 500 m; depth = 0 m
D dl»t. a. 1 m; d

1 mm/Min.

10 mm/Hr.

10 mm/Day

10 mm/Wk.

10 mm/Mo.

10 mm/Yr.

1 mm/Yr.

10 Yr. 1 Yr. 1 Mo. 1 Wk. 1 Day
I I I I I

*

1 Hr.

*

1 Min.

1 0-1-2 -3 -4 -5 -6

Log Time To Next Earthquake

Figure 14. The larger dynamic range of this log-log plot enables the increase in velocity
as the earthquake approaches to be more fully portrayed than does Figure 13. This figure also il-lustrates how premonitory displacements may serve.as the basis for an ideal earthquake prediction, following a criterion of Aki (1987) (see text). In order to provide an ideal prediction a linear relation would have to be found beteeen V/Vpl and probability of an earthquake occuring in the next minute, but such a relation seems at least qualitatively reasonable. Velocities are with respect to symmetry point at the fault as discussed in caption to Figure 7.

760

Velocity (meters per year)

0.04500 r-

0.04100 -

0.03600

0.03100 -

0 d«ptti = 0 mi Mm* = -207.24 doyt
+ d«pth a Om; Mm« = -173.88 day*
)£ d#pth = 0 m; tlm# = -87.78 doyt
Ddcptti « Om; Hm« = -24.13 day*

0.02600 -

2500 5000 7500 10000 12500 15000

Distance from fault (meters)

Figure 15. As the earthquake approaches, the velocities at the surface not only in-
crease everywhere, but the pattern changes so that the peak velocities occur away from the fault as a consequence of the increasing slip rate at depth on the fault. By a day before the earthquake the peak velocity has moved further, to a distance of 4.5 km from the fault. The pattern becomes progressively accentuated as shown in Figures 13 and 14. Beginning a few minutes before the earthquake, the peak in velocity at the surface shifts further to 6.5 km from the fault (as the point of highest velocity on the fault at depth rapidly moves to 6.5 km depth) and then jumps to the fault itself as the dynamic rupture occurs. Velocities are with respect to symmetry point at the fault as discussed in caption to Figure 7.

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Some Observations and Some Comments on Intermediate Term Precursors

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Many different types of intermediate term precursors have been reported, the most common being changes in the local seismicity. Recently an increasing number of strain measuring devices have been installed to monitor changes in the strain field. Although many of the following comments may have general applicability, the data discussed here is almost solely from borehole strainmeters. These comments are of necessity based on minimal data since there are not as yet a sufficient number of case histories from which to learn the typical (if any) strain changes that occur in the months to years before larger earthquakes. In addition, the various noise sources in this period band have not as yet been adequately studied.

Noise.

At periods less than a day, there are signals such as earth tides, surface waves and the like which allow the instrument fidelity to be evaluated. At longer periods there is the possibility that measurements reflect non-tectonic influences, such as rainfall, atmospheric pressure changes, changes in the aquifer system, etc. If the surrounding rock is not dimensionally stable (due for example, to mechanical instabilities such as fractures or to chemical changes such as in clay bearing minerals in the varying presence of water) this will also cause an instrument nearby to register strain changes which are due to small scale local effects and are not related to the signals of interest. In installations in fractured rock, there is always the possibility that there will be motion on a nearby fracture which will cause a large signal.
Sensitivity to rainfall has traditionally been a major difficulty with deformation measurements made at the surface. The shallow aquifers are recharged fairly rapidly after rain (over a period of hours to days). These aquifers then behave as hydraulic rams and produce relatively large strains in their vicinity. Figure 1 shows data from an extensometer in a tunnel which indicates a recharge time constant of about 1 day and a discharge time constant of about 3 days, i.e. pressure due to the rainfall episodes shown leaks away in about 3 days. For deeper installations, because of the low-pass filter characteristic of aquifers, each rainfall may not register but the season will, i.e. there can be a strong annual cycle in the strain data, see figure 2. The pressure in the aquifer is also changed by tectonic strains. Unfortunately, these pressures will leak away just as pressure due to rainfall does, and the data from a tectonic signal is effectively high-pass filtered. How far down in level the longer period signal falls depends on the rigidity of the rock matrix. Examples of signal attenuation in instruments with large annual cycles are shown and discussed in Linde and Sacks, 1987. While aquifer problems affect only a minority of installations, they can be important in the intermediate term precursor period band and should be evaluated for each site.

Precursor Signals.
Stress redistribution in earthquake zones can be complex, and non-linear in space and time (e.g. Sacks et al., 1978). In many regions one cannot consider a fault as an

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isolated system, subject only to steady loading and episodic release. In some cases con-centration of stress on a fault may be due to aseismic slip (or slow earthquakes) on ad-jacent orthogonal faults. In this situation, genuine precursors may occur substantially away from the eventual earthquake location. It is also possible that the locking of a fault system prior to an earthquake will cause stress redistribution in the surrounding faults which can cause detectable strain changes at these faults.
Ishibashi (this volume) has named this class of events "tectonic" precursors to differentiate them from "physical" precursors, i.e. those occurring in the preparation zone of the earthquake. With only a sparse network of instruments able to record these signals, it is difficult to determine with confidence which you have. This is further complicated by a possibly complex slip function on whichever fault is yielding. Figure 3 is such an example. It shows six months of data from two different types
of borehole strainmeters, spaced about 7 km apart. The large strain increase in early October is the co-seismic strain change due to the Morgan Hill (m = 6.5) earthquake. The instruments had only recently been installed and the large initial drifts have been removed. The similarity of the waveforms suggest that they are responding to the same strain change. It is beyond the scope of this discussion to interpret these signals in terms of a model, rather we note that
a) the strain field changes are coherent over a substantial distance (7 km).
b) There are many sub-events: 1) A step-wise decrease about.2 weeks before the earth-quake; 2) a gradual further increase for 2 weeks following the quake; 3) a gradual de-crease for about 3 weeks; 4) a rapid and substantial decrease for the next 2 months. In the last month shown the two sites are presumably responding to other, more local, changes. Note that if we were not fortunate enough to have two instruments within range, it would be impossible to estimate whether these events were in fact due to con-tinuing (and precursory) substantial motion on the Morgan Hill and adjacent faults, and therefore significant, or due to small slip events on a micro-fault very near an in-strument. In the latter case, if the source is very near, the strain changes recorded may have no tectonic significance. However, in the common situation of sparse arrays, we cannot disregard strain signals which appear on one instrument only. To summarize, we have the twin complications that slow strain changes may be complex, and that there may be motion on adjacent faults.

Examples.
In the first example, shown in figure 4, the precursors are clearly of the "tectonic" type. Slow strain changes were recorded by two instruments of different type for about five months prior to the Yamanashi earthquake, m = 5.5. Both the borehole dilatometer at Shizuoka and the extensometer at Fujiwara are within a few km and on opposite sides of the Fossa Magna (aseismic) fault. The polarity of strain change is opposite on the two instruments. The events terminated shortly after the earthquake. Since the events did not occur at the same time on the two stations, they may indicate strain propagation up the fault.
In the second example, because of the sparse instrument coverage, it is not yet clear whether the precursors were "physical" or not. The Gojonomae instrument of the Earthquake Prediction Institute of Tohoku University was operational about 7 months before the Japan Sea earthquake of 1984 with magnitude 7.8. Figure 5 is a calendar of bay-type events which were recognized by K. Suyehiro. The strain events had an amplitude of up to 3 x 10~8 and durations of a few hours. More of these events were recorded each successive month, there being 7 in the month preceding the quake. After the earthquake and its largest aftershocks, these anomalous events ceased. Until further

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modelling is completed, we cannot tell whether these events were in the preparation zone of the earthquake, some 60 km distant, or have been due to sympathetic strain changes on a fault nearer the instrument.
In the final example, we discuss the strain changes which occurred in the 1977 - 1978 period on the Izu peninsula of southern Honshu, Japan. The data came from a net of then 10 instruments (now over 30) which were installed by and are operated by the Japan Meteorological Agency (JMA). These data are published in the Reports of the Coordinating Committee for Earthquake Prediction and in the bulletins of the JMA. The first 5 instruments became operational in April 1976 and the second 5 one year later. Figure 6 shows the sites relevant to the discussion.
One of the most convincing precursors, in that it was recognized before the 1978 Izu-Oshima earthquake (m = 7), was recorded on the Irozaki dilatometer of the Japan Meteorological Agency network at the southern tip of the Izu peninsula, as well as at the Radon monitoring well in east-central Izu. Some water wells also showed anomalous levels during the 6 week period preceding the quake, but with a poorer signal to noise ratio. Since the form and timing of the anomaly is similar on the dilatometer and the Radon monitor, the strain change must be substantial and cannot be local to either instrument since they are about 40 km apart. The waveform of the anomaly on the strainmeter is shown in figure 7. Note from the map in figure 6, that there is one other instrument which is within range of the anomaly source, Ajiro on the east coast of the Izu peninsula. However, there is no signal at the onset time of the IRO signal, see figure 7, but a major trend change starts about 3 weeks later. We assume that AJI is on a nodal plane which would place the signal source either in the slow earthquake region on land (described in Sacks et al., 1981) or at the coastal termination of the Izu-Oshima marine quake. These regions are shown in figure 6. While the modelling of this source is beyond the scope of this paper, it is clear that it cannot be in the preparation zone of the earthquake, which is near Oshima island.

Discussion.

Some of the questions raised in the above examples can be answered after more complete modelling. However, the purpose of this paper is rather to indicate the variety and range of the intermediate term precursory strain changes which have been recorded in an active seismic region. Eventually we have to understand and interpret these signals. Until then, we need to maintain a rather open mind as to what "convincing" precursory signals will be.

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Figure Captions

Figure 1. Extensometer data. Strain resulting from rainfall. The rain starting on day 8 eventually gets into the aquifer (just before day 9) causing a local strain buildup over the next day, even though the rainfall is diminished. The strain change is about 60 times greater than would be expected from water loading of a uniform elastic medium. These data were kindly provided by the staff of the Imaide-yama Observatory.

Figure 2. Strain record from Mikkabi site showing clear annual cycle due to seasonal recharge of the aquifer.

Figure 3. Detrended strain recorded at San Juan Bautista (borehole tensor strainmeter, Gladwin 1984) and (upper trace) at Searles Road (dilatometer). Ordinate units are microstrain and time marks are months. The sub-event numbers are discussed in the text.

Figure 4. Time history of the slow earthquakes (recorded on the borehole strainmeter at Shizuoka and on the Fukigawa extensometers) preceding the m = 5.5 Yamanashi earthquake. The Shizuoka observatory began operation in April 1976. The amplitude scales for the two instruments are not the same, but the relative polarity of the strain steps is as indicated. Also included is a map showing the relative locations of the two observatories, the epicentre (O) and the Fossa Magna.

Figure 5. Calendar of pulses preceding Japan Sea earthquake of 1984. Pulses cease after the m = 6.9 aftershock.

Figure 6a. Map of southern-central Honshu, Japan, showing the borehole strainmeter sites discussed, IRO and AJI. 2 is the Izu-Oshima earthquake (m = 7, 14 Jan 78), 3 and 4 are later slow earthquakes identified by Sacks and Linde 1981. A precursor originating at the open circle, near the coastal termination of the Izu-Oshima earthquake would have a nodal plane through AJI, have compression at IRO and expansion at the Radon and water well sites, in agreement with observation.
Figure 6b. Precursor on Izu, at the position of the slow earthquake, also giving
a nodal plane at AJI. Broken lines indicate compression. The water well data is still satisfied, but IRO is in the expansion quadrant.

Figure 7. Strain data showing precursory changes on Irozaki and Ajiro prior to the Izu-Oshima earthquake, m = 7. See map on figure 6 for relevant locations. Note that Ajiro excursion starts later than that on Irozaki.

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Earthquake

Fujigawa extensometer
events

Fossa
moana

50km

Shizuoka starts
operation

January February March Apri

FIGURE 4

May June

On the Recognition of Intermediate-Term Crustal Deformation Precursors

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ABSTRACT

Most intermediate-term crustal deformation precursors reported in the literature have not been presented with sufficient background data (such as noise levels, false alarm rates and meteorological parameters) that their validity can be reliably assessed. Few, if any, con-vincing precursors of this type have been observed in the United States. Many of the methods currently used to search for these precursors are already operating close to their noise level. The past decade of reaearch has shown that short-baseline, near-surface instru-
ments are unsuited to detecting sub-microradian intermediate to long-term signals. Poten-
tially, the biggest improvements in signal detectability will come from installation of
sufficient numbers of lower-noise continuously recording instruments. Specifically, more
and better deep borehole strain and tilt meters, and long baseline tiltmeters, are required.

1. Introduction

It is difficult to provide a truly critical review of crustal deformation and creep precursors to earthquakes because of the lack of uniformity with which possible precursors have been reported in the literature. Presentations of precursors frequently consist of little more than the anomalous section of data, whereas for proper appreciation of the data, two additional items are required: (1) a comparison with historical data from the same instrument in order that noise levels and false-alarm rates might be assessed; (2) a comparison with meteorological data, since much of the observed signal in many crustal deformation records is undoubtedly due to weather. An additional problem is that before many earth-quakes with reported precursors, the precursors are observed at only one or a very few sites. With these provisos, I have reviewed many of the intermediate-term crustal deformation (hereafter ITCD) precursors that have been reported in the literature, and have attempted to reach some conclusions regarding their validity. I also propose techniques for detecting this class of precursor in the future, with particular regard to the situation in California.

Many compilations of precursory signals exist, most notably in works by Rikitake (e.g. Rikitake, 1976). References to these may be found in Bilham (1981); a more recent study is that by Mogi (1985). I have used these compilations as well as more recent sources (including Rikitake, 1978; Kiss-linger and Suzuki, 1982; Asada, 1982; UNDRO, 1983; UNESCO, 1984) to generate what I believe to be a fairly complete b'st of reported ITCD precursors and possible precursors (Table 1). Precursors have been reported from frequently repeated level lines, trilateration and triangulation, sea-level meas-urements, strainmeters, tiltmeters, creepmeters and gravimeters. Though water well measurements may also be considered crustal deformation observations, we do not treat them here because they are reviewed elsewhere at this meeting. The precursors generally take the form of a change in rate or direction of the quantity being measured, or they appear as an offset from the normal trend of the data that may or may not recover before the earthquake.

Most crustal deformation precursors fall into the intermediate- or short-term categories. Geodetic and sea-level measurements may also be useful as long-term precursors when they give information such as the position within the seismic cycle in a particular area; an intermediate-term precursor in this

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case might be a deviation from the long-term rate. The division between intermediate- and short-term crustal deformation precursors is less clear, and in fact many of the precursors that have been reported are short-term. We arbitrarily use a cut-off of 1 month for the precursors discussed in this paper (as suggested by the conference organisers).

2. The data

Rikitake (1976, 1979) studied many types of precursor and found a linear relation between log (precursor lead time) and magnitude (Fig. 1). It is notable that he omits instrumental strain and tilt pre-cursor data from these compilations because they do not fit well in his data set. Bilham (1981) attempted to make sense of these precursors, but failed to find any clear relation between anomaly size, lead time, distance and earthquake magnitude. Whether this is because many of the anomalies are illusory, or because the effects are real yet truly have no consistent onset time or magnitude, is unclear.

Many models of earthquake generation predict some sort of crustal deformation precursor. The dilatancy (Nur, 1972; Scholz et al., 1973; Whitcomb et al., 1973) and dry-dilatancy (Stuart, 1974; Mjachkin et al., 1975) models require uplift in the focal area, and models involving pre-slip on the eventual fault plane (e.g. Stuart, 1979; Stuart and Mavko, 1979; Rice and Rudnicki, 1979; Das and Scholz, 1981; Li and Rice, 1983a, 1983b) require accelerated, or occasionally reverse, deformation prior to the eventual rupture. The predicted signals are usually small, and therefore very quiet instruments will be necessary to observe them. As most recently discussed by Johnston et al. (1986), it is presently uncertain whether precursory anomalies appear over the whole of the future rupture zone, or whether they are localised at asperities. In the latter case they will be harder to detect since the affected volume is smaller.

Most reported precursors do not fit easily to any of the standard models. Indeed, Chinese, Japanese and Russian data suggest that the observed signals are often neither as small nor as localised as predicted. Mogi (1981) has suggested that the appearance of precursors is directly related to the heterogeneity, complexity and scale of the local tectonics (Fig. 2). Thus, earthquakes in parts of China have very many and widespread precursors, while they are more localised in Japan and almost non-existent in California. Bilham and Beavan (1979), and others dating back to Imamura (1933), have sug-gested that motions of crustal blocks in response to the build up of stress, are responsible for the distant precursor observations. This is suggested in particular by the Chinese experience, where large anomalies are frequently observed at certain "singular" points - usually the intersections of large frac-ture zones (e.g. Ma et al., 1982). In the U.S., possible premonitory signals on block boundaries have been observed by Leary and Malin (1984).

Even in cases where precursory signals have been discussed for many years in the literature, a closer examination of the data often reveals questions about the reality of the precursors. A good example is the 1964 Niigata earthquake (Fig. 3), where anomalies in leveling data (Dambara, 1973), sea level (Tsubokawa et al., 1964), tilt (Kasahara, 1964) and strain (Press and Brace, 1966) were all reported. Since then, the strain anomaly has been shown to be correlated with rainfall (Tsumura and Ando, 1972), the leveling anomaly has been partly explained as due to errors in the 1955 survey (e.g. Mogi, 1984), and the sea level anomaly is not apparent in more recent data analysis (Kato, 1983). An anomaly does appear to persist in the leveling data (Fig. 3b) but it is of long-term rather than intermediate-term character.

Nevertheless, the intermediate term precursors observed in some cases are sufficiently intriguing that it is necessary to persevere in trying to measure them. For example, for a year or so before the 1976 Tangshan earthquake, changes were observed on fault-crossing level lines, tiltmeters and strainme-ters, as well as by leveling and gravity surveys and by other (non-deformation) means. Though the reality of some of these signals is doubted by some workers, they are sufficiently persuasive to prompt further work.

3. Noise in the data

The noise problem for near-surface instruments has been studied in some detail for both temporal and spatial noise. A close relationship (Fig. 4) between strain and tilt signals and meteorological parameters such as rain, temperature and atmospheric pressure has been demonstrated in many instances

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TABLE 1

Possible intermediate-term crustal deformation precursors Date
1923

1933
1944
1945
1946
1952
1952
1952
1956
1960
1961
1961
1961
1961
1962
1963
1964

1966
1966
1967
1967
1969

1970
1970
1971

1971
1973

1974
1974
1975

1975
1976
1976
1976

1976
1977
1977
1978

1978
1978
1978
1979
1983

1983
1984

Event
Kanto

Long Beach
Tonankai
Mikawa
Nankai
Kern Co.

Daishoji-oki
Yoshino
Dunaharaszti

Odaigahara
Nagaoka
Hyuganada
Kitamino
Hollister
Shirahama
Echizenmisaki
Niigata

Paikfield
Peru
Luchobsk
Koyna
Gifu

Akita
Tien Shan
Atsumi-oki

San Fernando
Nemuro

Hollister
Izu

Haicheng

Kalapana
Wellington
Gazli
Tangshan

Longling
Briones Hills
Wellington
Izu-Oshima

Tien Shan
Tabas-e-Golshan

Alaysky
Coyote Lake
Japan Sea

E. Yamanashi
San Juan

M
7.9

6.3
8.0
7.1

7.2
6.8
7.0
6
6.0
5.2
7.0
7.0

6.4
6.9
7.5

7.9
4.5
7
6.6

6.2
6.7
6.1

6.4
7.7

5.2
6.9
7.3

7.2
5.5
7.3
7.8

7.4
4.3
6.2
7.0

6.6
7.7
7.0
5.9
7.7

6.0
3.1

Type
Triangulation
Sea level
Tilt
Sea level

Leveling
Sea level
Sea level
Short leveling
Tilt
Strain

Leveling
Tilt

Leveling
Sea level

Creep increase

Sea level
Tilt
Tilt
Creep (visual)
Sea level
Strain

Tidal admitt.
Tidal admitt.
Strain
Strain
Tilt
Short leveling
Tilt
Strain

Leveling
Tilt
Strain
Tilt
Strain

Short leveling
Tilt
Gravity
Trilateration
Tilt
Tilt
Short leveling
Short leveling
Tilt
Gravity
Leveling
Short leveling
Tilt
Tilt
Volume strain
Volume strain
Tilt
Tilt

Leveling
Creep retard.
Tilt
Sea level
Tilt
Shear strain

Station
Mitaka
Aburatsubo
Tokyo

Kakegawa

Uwajima

Ogoya
Osakayama

Kii peninsula

Nezugaseki
Maze
Maze
Parkfield

Kondara

Kamitakara
Inuyama
Kamitakara
Inuyama
Nibetsu
Alma Ata
Inuyama
Inuyama
Saugus-Palmdale
Erimo
Erimo
Hollister
Fujikawa
Jin Xian

Shenyang

Khandallah
Tashkent
Dahuichang
Niukouyu
Dahuichang
Tianjin
Ninghe
Xiaquan
Berkeley
Khandallah
Irozaki
Ajiro
Turgen
Ashkabad
Andizhan
Hollister
Oga
Fukawa, Oga
Enzan
San Juan

Lead time

4 yr
10 yr
1 mo

4-5 yr
1 mo

4-5 yr
Syr
5-10 yr
3 mo
10 mo

lyr
6 mo

3yr
4-5 yr
1-2 mo
2-3 yr
1 mo
1-6 mo

lyr
Syr
6 mo
weeks?
5-6 yr
6 mo
3-10 mo
8-9 mo
8 mo
8-9 mo
8 mo
2 mo
2-3 yr
8 mo
8 mo
years
6 mo
1-2 mo
1 mo
7 mo

1-2 yr
2 mo

3yr
4 yr
1-2 mo

lyr
lyr
lyr
lyr
3-4 yr
2-3 mo

2yr
1 mo
7 mo
1-2 mo
1 mo
2-3 mo

4 yr
1 mo

3yr
4-5 yr
Syr
<1 mo
1 mo

(e.g. Rikitake, 1976; Wood and King, 1977; Kasahara et al., 1983; Ben-Zion and Leary, 1986). Noise due to variations in groundwater level has also been documented (e.g. Kumpel, 1983; Evans and Wyatt, 1984). A few studies shed light on the length of instrument or depth of burial that are required to sufficiently isolate the instrument from such effects. Wyatt et al. (1982) compare short (1m) and long (500 m) baseline tiltmeters and show that the latter are 1-2 orders of magnitude quieter at all but tidal

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periods (Fig. 5). Evans and Wyatt (1984) show that near surface cracks can have a profound effect on signals measured near the surface (Fig. 6). Extending this work, Evans (pers. comm., 1986) suggests that joints and fractures have developed substantial normal stiffness at stress levels anticipated for depths of 200 m., assuming a normal hydrostatic and lithostatic stress state (Fig. 7). A depth of 200 m should therefore be a minimum target depth for emplacement of borehole tilt and strain instrumentation. Evans et al. (1979) show that strain amplitudes measured by an array of 10 m instruments separated by hundreds of m can vary by as much as 50% (Fig. 8). This suggests that lengths of near-surface instru-ments should be least hundreds of metres in order to average over local inhomogeneities. Savage et al. (1979) find similar results in a study of short level line data.

4. Some case studies

One of the more convincing precursors is that measured by trilateration before the 1975 Kalapana, Hawaii earthquake (Fig. 9) (Wyss et al., 1981). It is convincing because it is associated with other anomalous parameters, in this case p-velocity variations and seismicity rates and patterns. It is also one of the few examples that fits well with a model - in this case strain softening. Unfortunately, the tec-tonic stresses that gave rise to the earthquake resulted from magma injection, so that loading rates were an order of magnitude greater than in typical earthquake situations. Similar effects might not be so apparent at slower loading rates.

In order to emphasise the necessity for collecting and presenting long-term data, and for monitor-
ing environmental parameters, I discuss three examples. The first is the 1977 Briones Hills swarm (Fig.
10) (Johnston et al., 1978). Fig. lOa shows the anomaly in the dramatic form in which it has some-times been presented in the literature, even though the original authors were far more circumspect in its interpretation, and though independent deformation data from the same area showed no anomaly (Savage and Prescott, 1978). Much of the anomalous signal is probably due to the December 30 rain-fall, and if post-Dec. 30 data are omitted it becomes impossible a priori to recognise any unusual change. The amplitude of the signal is marginally, if at all, bigger than other signals in the previous three years of data from that instrument (Fig. lOb), and is smaller than signals measured on the other tilt component at the same site. Another example of postulated precursors being presented in isolation are the volume strain anomalies detected prior to the 1978 Izu-Oshima earthquake. These have been published in the suggestive form shown in Fig. 11 a, whereas a plot of the entire data set (Fig. lib) casts the anomalies in an entirely different light (though without necessarily invalidating them). Incidentally, this is another example of anomalies that may be due to magma injection (e.g. Linde et al., 1983; Thatcher and Savage, 1983). The third example (Fig. 12) is a precursor with rather less than 1 month lead time, but I include it as it illustrates the type of data collection and presentation that is necessary to reliably detect precursors. An anomalous tilt change of 0.4 |irad occurred in 10 days, starting 18 days before the 1983 East Yamanashi earthquake (Sato et al., 1984; Sato, 1986). This tilt was not rainfall related, and is substantially larger than any other non-typhoon related signal in the pre-vious 4 years of record (Fig. 12b).

An increasing number of researchers have been reporting negative results, as well as positive ones, in searches for precursors. This is useful as it enables a better data base to be constructed on noise levels and false alarms. Recent examples are the 1983 Coalinga, 1984 Morgan Hills, 1984 Round Valley, and 1985 Kettleman Hills earthquakes (Johnston et al., 1986; Galehouse, 1986), none of which showed precursory anomalies on nearby volume strainmeters or triangulation nets. These authors were, however, concentrating on short-term, rather than intermediate-term precursors.

5. Crustal Deformation in California

There are two possibilities for the lack of convincing precursory observations in California. One
is that precursors do not occur. The second is that they have not occurred at the noise level of existing instrumentation. (By "instrumentation", I mean to include geodetic observations as well as instrumental measurements.) Based on theoretical models, and on results from elsewhere, we believe that the latter is the case. Our approach must therefore be to deploy quieter instruments. (The alternative is to wait for larger earthquakes in the hope, by no means proven, that their anomalies will be larger). There is ample evidence for present day deformation in California from geodimeter networks, VLBI and SLR

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measurements, multiwavelength distance measurements, leveling, etc. The problem is to detect anomalies in these data that might signify approaching earthquakes. It is even possible that some sig-nals already detected, such as the deflated (and perhaps illusory) Palmdale bulge, may prove to be pre-cursors of events yet to come. An interesting correlation between strain, gravity and elevation changes was noted over southern California by Jachens et al. (1983). This correlation appears to continue through the present (Jachens, pers. comm., 1986; Johnston, 1986) with an additional correlation to mag-netic data. However, Savage et al. (1986) find little evidence in the strain (trilateration) data alone for anything other than linear strain accumulation, so the correlated changes cannot be regarded as proven. The variations in these data sets perhaps show the typical noise levels that can be expected from such data, and above which precursors must appear if they are to be recognised.

Creep retardation (Schultz et al., 1982, 1983) appears a possible precursor on the central section
of the San Andreas fault system, particularly in view of its possible relationship with seismic quiescence (Burford and Schultz, 1985; Wyss and Burford, 1985). This may be of great importance for central California quakes, but it appears to have little application elsewhere.

6. Strategies for the future

In order to be useful for public policy and planning purposes, it is essential that precursors be clearly identified. To me, this means that any ITCD precursor should have clearly occurred at more than one site, and should preferably be accompanied by precursors in at least one other type of meas-urement. A prerequisite for this is that instruments should be as quiet as possible, should run reliably for long periods so that noise levels can be assessed, and should be accompanied by environmental
measurements.

There are three strategies for reducing noise in crustal deformation measurements.

(1) make the instrument big: e.g. traditional geodesy, space geodesy, multi-wavelength
geodesy with fixed central station, sea level, lake level, water-tube tilt, laser strain. (Tradi-
tional geodesy and space geodesy measurements are often too infrequent for useful intermediate-term data, but in areas where a long-term forecast has been made they can be intensified.)

(2) locate the instrument deep: e.g. deep borehole tiltmeter, strainmeter.

(3) use natural amplification (if it can be demonstrated that signal: noise ratio can be increased): e.g. strainmeters across cracks or faults.

I believe that the geodetic measurements are in good shape. Savage et al.'s geodolite measure-ments are providing very useful data. They are probably operating near the limits of their accuracy, though more is continuing to be learned about noise levels as more data are collected. Multi-wavelength measurements have not yet reached their ultimate accuracy, though here the problem is probably bench mark stability as much as anything else. GPS and VLBI measurements will continue to improve in density of observations and accuracy, and can be expected to make a major contribution over the next decade.

The problems lie more with the instrumental measurements. The advantage of these measure-ments is, of course, that they give a continuous, rather than sampled, history of any deformation that occurs. Differential sea level and lake level measurements are capable of giving low-noise data (better than 1 jirad, but not as good as 0.1 jirad at monthly to yearly periods). Their use should be extended, though their possible locations are limited. Most strain and tilt instruments do not have sufficient long-term stability that they could observe the longer (i.e. years) ITCD precursors. An exception is the long-baseline half-filled water tube tiltmeter with deep-reference end mounts (Wyatt et al., 1984). Cer-tain deep-borehole tiltmeters (e.g. Askania) have demonstrated good long-term stability; one is now under test with other tiltmeters at Pinon Flat (Wyatt, pers. comm., 1986). Borehole strainmeters do not appear in general to have good long-term stability, though they may already be useful for shorter (say, less than 1 year) precursors. For example, a 3-component borehole strainmeter (Gladwin et al., 1986) at Pinon Flat shows a shear strain rate several times higher than that measured geodetically. Despite this shortcoming (which may disappear with future improvements to the instrumentation), I believe the only way we are going to obtain convincing results is to make the instruments long or deep, to make them

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reliable enough to run for long periods of time, and to thoroughly test them before committing to large-scale deployment. The use of relatively cheap, near-surface instruments for sub-microradian inter-mediate to long term measurements, has been proved ineffective by the last decade or so of research. Borehole instruments that are removable would be advantageous for maintenance purposes. Present borehole strainmeters are all short baseline, and therefore have an inherent disadvantage in terms of sta-bility (i.e. very small instrumental and mounting instabilities appear as large apparent "regional" strains). Tiltmeters in cased boreholes have an advantage in that the modulus of the casing is typically an order of magnitude larger than that of the surrounding rock, and hence the casing will average tilt over a larger baseline than the instrument itself. Perhaps future strainmeter development could take advantage of similar averaging.

An instrumental deployment and development program like this is necessarily expensive and long-term, and it will take a long-term, committed effort to provide a sufficiently dense instrumental coverage. It should be borne in mind that the principal expense of operating an instrument that will run for decades is the routine maintenance and data gathering; therefore, maximum effort should be made to ensure the quality of the initial installation.

7. Conclusions

Most reported ITCD precursors are not presented with enough back-up data that their validity can
be assessed. Few, if any, convincing ITCD precursors have been reported from the U.S. mainland. The most likely way to rectify this situation is to instal high quality, low noise, instruments in sufficiently dense networks (as is being attempted at Parkfield). There is some hope that near-surface measure-ments within fault zones can be used to detect precursory signals, and this should be followed up on a small scale in order to determine whether signal-to-noise ratio is actually increased in such zones.. My recommendations for the future earthquake-prediction-related measurement program in California are as follows:

1. Continue and expand geodetic work
2. Stop any work that is still continuing on shallow, short-baseline instruments
3. Continue installation of and improvements to deep borehole and long-baseline surface instruments

4. Continue sea-level and lake-level measurements, and expand them in the limited regions
where they are applicable
5. Continue creep measurements in appropriate places; improve creep instrumentation
6. Small scale investigation of measurements across fault zones

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Logarithmic precursor time in days versus magnitude plots for the
precursor data excluding those for disciplines t and f. Each alphabet indicates
a discipline as listed in Table II (RIKITAKE, 1978b).

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Figure 1. Rikitake's relationship between precursor lead time and earthquake magnitude. Instrumental strain and tilt precursors are not included (from Rikitake, 1979).

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Simplified tectonics and the nature of precursory phenomena in central California, Japan and China.

Figure 2. Mogi's suggestion for the differing occurrence of precursors in the U.S., Japan and China
(from Mogi, 1981).

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Changes in height at levelling bench marks before and after the
1964 Niigata earthquake (DAMBARA, 1973).

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Figure 3. (a) Niigata leveling precursor as presented by Rikitake, 1976; (b) same, as modified by Mogi (plot reproduced from Mogi, 1984).

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Correlation between strain and rainfall at five points in the Tokai district, June-September 1977.
Distances in parentheses are the depths of the embedded instruments.

CONTINUOUS OBSERVATION OF CRUSTAL MOVEMENT 155

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Correlation between strain and atmospheric pressure at five points in the Tokai district, Feb. 9-28, 1977.

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Figure 4. Atmospheric noise signals in strain and tilt data. Correlations with (a) atmospheric pressure (from Rikitake, 1976); (b) rainfall (Rikitake, 1976).

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Paso Robles

These data are from a variety of
geophysical instruments implanted in the soil near to or within the central California section

of the San Andreas fault system. Maximum separation of sensors is approximately 275 km. The time bracketed by the dashed lines is the only period of rainfall for the time of the record shown (5). The abbreviations XFLl, XMR1, and so forth designate site names.

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Multiple time series starting with (curve 1) observed daily tilt from Libby-N, (curve 2) tilt predicted from rainfall, (curve 3) residual (difference) between curves 2 and. 1 , (curve 4) residual predicted from temperature, and (curve 5) final residual, that is, curve 4 minus curve 3. Occur-rence time of largest earthquake (M » 5.2) on 28 November 1974 is denoted by *.

Figure 4. Atmospheric noise signals in strain and tilt data. Correlations with (c) rainfall (Wood and King, 1977); (d) rainfall and temperature (Wood and King, 1977).

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TILT POWER SPECTRA

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Frequency (cpd)
Power spectra and relative coherence for the tilt-meter records in Figures 1 and 3. Only the UCSD and LDGO tiltmeters show significant coherence (y 2 ^ 0.18) outside the tidal bands.

Figure 5. Noise spectra of shallow borehole versus long-baseline tiltmeters (Wyatt et al., 1982).

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Fig. 4. Same as Fig. 3. with an estimate of the earth tides removed to better reveal the drilling-related disturbance. Approximate duration of downliole activity for days 299-302 were 4.75, S.2S. 4.S and 4.2 hrs respectively.

Figure 6. Surface displacement signals in response to variations in ground-water pressure (Evans and
Wyatt, 1984).

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Figure 7. The effect of near surface cracks on elastic moduli (K.F. Evans, pers. comm., 1986). (a) and (b) show fracture density and p velocity as a function of depth in the Crystallaire hole. The approach of in-situ velocity to the laboratory values is shown, (c) shows two laboratory determinations of modulus"1 for granite, (d) shows the effective modulus"1 due to fractures alone, as derived from data of Bandis et al. (1983). The effect of fractures is substantially diminished below 200 m. Pore pressure is ignored in this figure; however, the figure suggests that cracks will have a serious effect whenever hydrostatic pressure approaches within ~ 60 bars of lithostatic.

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Observed tidal admittances at three of the linear array sites. At QNB both M 3 and O, results are shown; at WDH and HCS only the M, tide was resolved on all instruments. The top of each diagram shows the topography with no vertical exaggeration, and the positions of the strainmeters together with their associated data numbers. At QNB the strainmeter identification numbers are also shown because in two cases the same strainmeter was installed at different locations in the tunnel. The bottom of each diagram .shows the distance in metres from the tunnel entrance. \r \ and 0 are the modulus and phase lag of the admittance of the observed tide with respect to the theoretical solid Earth tide in the same latitude, longitude and azimuth. The horizontal dashed lines show the values of \r\ and 0 predicted from calculations of the ocean load tide. (For O1 the ocean load is assumed zero.) The data are shown with error, bars at the 95 per cent confidence level (if dashed, no estimate available). Note that the QNB results are on a larger scale than the others. The greater variations in the WDH and HCS data are probably due to their much higher noise levels.

Figure 8. Effects of local inhomogeneity on strain amplitudes (Evans et al., 1979).

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TIME (YEARS)

Schematic summary of crustal parameter changes during the 8 years before the Kalapana earthquake as a function of time. Arrows at the top mark the beginning of eruptions (Table 1). Hie geodetically measured strain changes are ihown as the two top curves for the inner and outer anomalous areas separately. The second curve shows the relative P residuals observed in region 2 as a solid curve and the constant P resi-duals observed in the outer anomalous volume as a dashed line. The third curve shows the seismi-city rate of the outer region (solid line) and that for the two inner anomalous regions (dashed line) normalized to 100%. The fourth curve shows the average epicentral distance from the M = 5.9 foreshock which measures seismicity bunching toward the approaching main rupture. The bot-tom curve shows b values (average magnitude) as one curve for the outer and inner regions.

Figure 9. Precursors of the 1975 Kalapana earthquake (Wyss et al. 1981).

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Expanded record of the tilt time history on BK8 round the time of the Briones earth-
quake swarm. Earthquakes and rainfall are marked as for Figure 2a. Least-squares quadratic
fits to the apparent smooth trend prior to December 20 and after January 24 are shown with dashed lines.

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Tilt recorded at BKR for 3 years from January 1, 197-1, to March 1977. Shown also is
rainfall recordt d nt the Berkeley Weather Station and t no occurrence time of earthquakes (arrows)
with ML > 2.5 and epicentral distance <10 km.

Figure 10. (a) Briones Hills tilt anomaly as often presented; (b) 3 years data from the same instru-ments (Johnston ct al., 1978).

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Omaezaki Irozaki

Observed result made by the volumetric strainmeters of the Sacks-Evertson type at fou: stations around the 1978 Izu-Oshima-kinkai earthquake (JMA, 1978).

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TIME (DAYS)

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Strain records from AJI and IRQ on the Izu peninsula and from IRA, which is to the west. Compression is positive. Note tin: AJI is scaled by a factor of 0.3 for comparison purposes. The long dashed lines indicate the 14 January 1978 and 29 June 1980 earthquakes. The shortest dashed line indicates
an m - 4.1 earthquake about 20 km from AJI. Small vertical ticks near the time axis indicate small earthquakes within 25 km of AJI. The longer solid vertical lines denote the occurrence of swarms within
25 km of AJI. The solic circles (at times indicated by A, B, and O denote the strain changes calculated
from the magma model (see text. Fig. 41, Fig. 42). The first point on each curve is an arbitrary starting
value. Theoretical values were scaled by a factor of four.

Figure 11. (a) Izu-Oshima-kinkai strain anomaly as often presented (extension positive); (b) 5 years data from the same instruments (compression positive) (Linde et al., 1983).

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Hourly plots of tilt change at ENZ, AKW, ASG, and underground temperature
at ENZ, daily precipitation at ENZ, and daily number of earthquakes observed at
TRU (trace amplitude >64/«kine) for May 1-Aug. 31, 1983. The anomalous tilt changes arc surrounded by dotted curves.

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Figure 12. (a) East Yamanashi tilt anomaly; (b) 4 years data from the same instrument (Sato et al.,1984).

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Plots of daily means of tilt data at ENZ,
and daily precipitation at ENZ and KNM for Oct. 1979-Sep. 1983. Tilt steps of mechanical origin caused by earthquake shocks are cor-rected. Arrows indicate the occurrence of earthquakes of magnitude greater than 3 in the eastern Yamanashi active seismicity area, and symbol Ty means typhoon. The anomalous tilt changes are surrounded by dotted curves.

KINEMATICS AND MECHANICS OF TECTONIC BLOCK ROTATIONS: A REVIEW

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ABSTRACT

Large portions of the earth's crust are broken by dense sets of parallel or subparallel faults which are organized in domains. Recent analysis of structural and paleomagnetic data (Ron et al., 1984) suggest that when such domains are subject to tectonic shearing they deform by distributed fault slip and block rotations, rather than by uniform straining. Many such domains have been recognized in the Western U.S., and in California and Nevada in particular. Mechanical considerations (Nur et al., 1986) further suggest that when these rotations are sufficiently large (25" -45") new fault sets must form leading to domains of multiple sets. Several domains of multiple fault sets have by now been recognized in several localities in California and Nevada.

Because fault sets and domains are common in California and Nevada, we believe that block rotations must be common too and that the concept may be very useful in understanding past and present crustal deformation in the Western U.S.

By combining paleomagnetic data, structural geology, and rock mechanics we explore the validity of the block rotation concept and its significance in (1) Northern Israel, where fault slip and spacing is used to predict block rotation; (2) the Mojave Desert, with well-documented strike-slip fault sets, organized in at least three major domains. A new set of faults trending N-S is in the process of formation here; and (3) the Lake Mead, Nevada, fault system with well-defined sets of strike-slip faults, which, in contrast with the Mojave region, are surrounded with domains of normal faults; and (4) the San Gabriel Mountains domain, with a multiple set of strike-slip faults can be recognized.

Block rotations can have profound influence on the interpretation of geodetic measurements and the inversion of geodetic data, especially the type collected in GPS surveys. Furthermore, block rotations and domain boundaries may be involved in creating the heterogeneities along active fault systems which are responsible for the initiation and termination of earthquake rupture.

1. THE PROBLEM: DEFORMATION OF THE CRUST WITH DISTRIBUTED FAULT SETS

More than many other active plate boundary regions, Western North America has been investigated in great detail for decades. Due to the relatively dry climate, exceptional exposures of rocks and geologic structures exist in this region, exposures which attracted many geological and geophysical studies. The potential damage of future large earthquakes has attracted also a great deal of geophysical research on the distribution and activity of faults throughout this region. It is not surprising therefore that some major and classical geological and geophysical concepts of crustal deformation have developed here, including the idea of very large offsets on strike-slip faults (), the earthquake rebound theory (), crustal extension in the Basin and Range (), and the concepts of pull-apart basins (), accreted terranes (), and large crustal rotations ().

Despite this past and ongoing research activity much confusion still exists as to the details
of crustal deformation in the Western United States. This is due, more than anything, to the fact that much of the deformation here is distributed over a huge network of numerous large and small faults oriented in many different directions, and active more or less simultaneously. In this

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paper we review a new approach to the kinematics and mechanics of the deformation of crustal regions which contain distributed fault systems, with particular emphasis on selected portions of Western USA.

There are four specific features of distributed fault systems which are directly relevant to
the problem we wish to consider here: (1) the existence of sets of many faults, particularly sets of strike-slip faults which parallel one another; (2) the organization of these sets in domains; (3) the systematic tectonic rotations of block between faults within sets in a given domain; and (4) the presence within some domains of several intersecting fault sets, or multiple sets.

(a) Fault sets. Throughout the Western United States, and in just about any other tectonic regions in the world, sets of roughly parallel strike-slip faults are found (fig. la). These faults are active, within a set, roughly simultaneously. Although such sets are to be expected as part of major fault systems, e.g. the San Andreas system, sets are also common away from the San Andreas, distributed as they are throughout the Mojave Desert, and in large portions of the Basin and Range Province. Some questions which arise are: why do these sets exist, and what controls the spacing between faults within these sets?

(b) Domains of fault sets. Strike-slip faults tend to occur in domains of left- or right-slipping faults in the same tectonic environment, e.g. the three domains of the Mojave Desert (fig. 1). Unlike classic models (Billings, 1972; Anderson, 1951) which predict slip on conjugate faults, deformation actually seems to occur in "conjugate" domains where one domain may show left-lateral slip, and a neighboring domain may show right-lateral slip. This pattern has also been recognized in many distributed strike-slip systems worldwide. Why these domains form, what controls their dimensions and how are these related to slip and overall crustal deformation are some of the key questions still to be answered.

(c) Block and fault rotation. One of the most important aspects of fault sets in domains
is the rotation of the blocks between the faults when they slip, and the consequent rotation of the faults themselves (Freund, 1970a, 1974; Garfunkel, 1974). That these rotations take place has by now been established by comparing paleomagnetic with structural data (e.g. Ron et al., 1984) with surprisingly good agreement in areas where both data were obtained.

Simple considerations (Freund, 1970a) show that block and fault rotations associated with fault slip imply that sets with left-lateral slip undergo clockwise rotation, whereas right-lateral sets undergo counterclockwise rotation (fig. 2). These considerations suggest that fault sets and their domain geometry are time variable. An important question which arises here is: How do domain boundaries accommodate the differences in rotations between two adjacent domains?

(d) Multiple sets. Finally there is an important mechanical aspect of distributed fault sets
and their rotations. As faults in a set rotate away from the direction of optimal failure, they eventually must lock up in directions relative to the stress field in which slip is inhibited. Further deformation must then be accommodated on newly formed faults which are more favorably oriented to the stress. The creation of these new sets leads to domains with multiple sets of faults, with older sets offset by younger ones. Multiple sets have been recognized in several domains in Western North America and elsewhere (e.g. fig. 12). The obvious questions which arise are: under what conditions do new sets develop, and how are they related geometrically (length, spacing, termination) to the older sets?

2. THE MODEL: BLOCK AND FAULT ROTATIONS

During the past few years, we have (Ron et al., 1984; Nur et al., 1986; Ron et al., 1986) investigated two aspects of crustal deformation in regions with pervasive sets and multiple sets of strike-slip faults, as outlined in the previous section, by the combination of inputs from four separate fields: (1) structural and field geology; (2) paleomagnetism; (3) rock mechanics; and (4)

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seismicity. The synthesis of these inputs was based on a two-pronged analysis of the tectonics of regions with fault sets: (l) the kinematics, which involve the description of block and fault rotations, the description of fault sets and their domains, the geometrical and spatial relation between fault spacing, slip and directions of faults within sets and domains; and (2) the mechanics, which involve fault and block deformation and rotations subject to stresses and constrained by the frictional and fracture strength properties of crustal rocks.

Kinematics

It has long been recognized (Freund, 1970a) that fault blocks in strike-slip tectonic domains must progressively rotate on vertical axes as the overall strike-slip motion continues. Two direct consequences of this deformation mechanism are that (1) slip on each of the faults within a domain must be related to the rotation of the blocks bounded by these faults, and that (2) the faults themselves must also rotate because they are the boundaries of the blocks (fig. 2). These ideas were applied to strike-slip tectonics in Eastern Iran (Freund, 1970a), New Zealand (Freund, 1971), the Dead Sea transform (Garfunkel, 1970), Mojave Desert, California (Garfunkel, 1974), Southern California (Luyendyk et al., 1980), and Southeast Sinai (Frei, 1980). The actual geometric relations involved in this rotation process were analyzed by Freund (1974), Garfunkel (1974), and MacDonald (1980), who showed how the rotations of blocks and strike-slip displacement are two interrelated and contemporaneous aspects of a single deformation process. Specifically if the blocks are rigid, then the model predicts a simple quantitative relationship between fault spacing, slip, and amount and sense of block rotation. Thus in areas where fault spacing and net slip can be determined geologically, paleomagnetic measurements may be used to test the sense and amount of rotation predicted by the model. The simplest aspect of the rigid fault and block rotation model, Freund (1970a, b; 1974) and Garfunkel (1974) showed that the sense of block rotation must be opposite to the sense of the fault slip, with left-handed slip associated with clockwise rotation, and right-handed slip with counterclockwise rotation.

The geometrical relation (fig. 3) between the displacement d along a fault (positive when right-lateral), the width w of the faulted block, the initial angle a between the faults and the boundary of the domain, and block rotation 0 (positive when counterclockwise) is given by

j / sin<j> ti\ d w=- . ; (i)

The relative elongation /// 0 of the faulted domain parallel to its boundary, is given by
sina

The deformation within a fault domain is simple shear. However deformation resembling pure shear arises for example when two domains of conjugate faults bisect the angle between the faults.

The termination of faults at their domain boundary may be rather complex and is not well understood at present. In clay experiments, Hoeppener et al. (1969) found that faults tend to terminate by bending so that strike-slip motion can be transformed into dip-slip and/or diffuse strike deformation, e.g. by splaying (Freund, 1974).

The block rotation model can be rigorously tested in areas where domains of strike-slip faults exist and where the blocks are fairly rigid and where paleomagnetic declinations can be measured to reveal possible horizontal rotation about vertical axes. If the model is correct, we expect to find agreement using equation (l) between structural rotations computed from the amount and sense of fault slip and fault spacing as measured in the field, and rotation amount and sense obtained from paleomagnetic declination measurements (Ron et al., 1984).

Mechanics

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In many published cases, the rotations inferred from paleomagnetic data are large, exceeding 45"; some values approach 100" of block rotation (table l). If the faults rotate with the blocks, as envisioned by the block rotation model, then a major question arises: How much can a fault plane rotate away from the optimal direction of shearing before it cannot slip any more? If the magnitude of such rotation is restricted, then the kinematic model as described above must be modified accordingly.

An obvious constraint on the amount of fault rotation is imposed by the mechanical condition of faulting, namely that shear stress on the plane of the fault must exceed the shear resistance to slip along the fault. From extensive laboratory studies, it has been established that the shear stress r required for sliding on a fracture is controlled by the cohesive strength S lt and the normal effective stress <TO acting on the fracture.

T = S l + //a0 = 5, + n(a-p } (3)

where // is the coefficient of friction, a is the normal stress, and p is pore pressure. If we consider now a fault originally formed at the optimal direction of failure ^>Q relative to the maximum stress as shown in the Mohr diagram of figure 4, it follows that as deformation and rotation proceed, the shear stress acting on the fault plane decreases, and the normal stress increases. Beyond some critical angle of rotation (j> e it should become easier for a new fault to in the rock mass (with cohesive strength 5 0) form in direction <£0, relative to the stationary direction of maximum stress, rather than continue sliding on the preexisting one, now oriented in the direction <f> e from <£0.

The critical angle (f> e is given by (Nur et al., 1986)
- I

iJL tJ 1 / iJ Q I I

- - - - - (4)

J

where S Q and Si are the cohesive strength of the virgin rock mass and the preexisting fracture respectively, 0 = tan"1 //, and cr0 is overburden pressure.

Equation (4) shows that there is a mechanical limit on fault rotation under a stationary stress field. Apparently, the magnitude of the permissible rotation <£ c is dependent on the difference between the cohesive strength 5 0 of the virgin rock and that of the preexisting faults Si, the coefficient of friction // = tan0, and effective overburden pressure aQ (or equivalently, depth).

The results of equation (4), plotted in figure 5, show that the kinematic model must be modified, as shown in figure 6, to include the appearance of a new set of shear faults when rotation is sufficiently large. The values of <£ e obtained from equation (2) show that the angle to which a fault set can rotate before a new set must appear to accommodate further block rotation is in the range of 25* to 45° . For the extreme cases // = 0 or <TO = 0, we obtain the maximum possible angle of rotation of $e = 45" , which is the absolute upper limit of block rotation which can be accommodated by one set of faults. In situ, the angle <fr c should be the angle between the direction of the faults in the older locked set, and the orientation of the faults in the newer set, which offsets the older one.

3. FIELD EVIDENCE

Northern Israel. Probably the first suggestion that large spatial variation of paleomagnetic declinations may be due to tectonic rotations about vertical axes was made by Nur and Helsley (1971) as a result of a field study in Northern Israel in 1967. These results revealed large declination anomalies which have led Freund and Nur to propose a rigorous test of the block tectonics model. This test was carried out by Ron et al. (1984) in Northern Israel (fig. 7), because good structural data regarding fault spacing and slip were available and good paleomagnetic material yielded reliable and sufficiently accurate declination data.

The results of this study both structural and paleomagnetic revealed several domains of fault sets, some with clockwise and left-lateral fault slip, others with counterclockwise and right-

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lateral slip, as predicted by the model of Freund (1970). Figure 8 shows a summary of the structurally observed rotations vs. the paleomagnetically measured ones, with values of the rotations in the various domains ranging from 0" to 54°. The results clearly show that the magnitude and sense of block rotations as predicted from the structural data agree well with values obtained from the independent paleomagnetic determinations, suggesting that the rigid fault rotation process may thus be typical of such systems in general.

Multiple sets. What about multiple fault sets? There are a few in situ cases in which the angle <f> e between the directions of faults in an older and a newer fault set within one domain have been determined: The values of these angles range between 25° to 45°, well within the predicted range of mechanically permissible fault rotations. These values are, however, significantly lower than many of the total block rotations determined paleomagnetically (table 1). According to our model block rotations larger than 45" must involve two or more sets of rotating faults; and rotations over 90° require three sets or more. One of the examples of table 1 was recognized by Freund (1970) in the Sistan, Iran, region, where two sets of strike-slip faults 40° to 45" apart were found to have accommodated the rotation of crustal blocks. Three sets may be involved in north-central Iceland (Young et al., 1985), and in the Snake Range, Nevada (Gans and Miller, 1983). Although the rotation here is associated with normal, not strike-slip, faulting, the model is equally applicable. It is noteworthy that the angles between the sets as reported by Gans and Miller are again within the range of values permissible by the model. Proffett (1977) has similarly identified at least two multiple normal fault sets in the Yerington, Nevada, area with an angle between sets of around 40" .

The applicability and usefulness of the block tectonics approach in the Western U.S. can be evaluated by considering two areas the Mojave Desert, and the Lake Mead-Nevada fault zone-where comparisons between structurally inferred rotations, as derived from fault geometry and slip, and paleomagnetically derived rotations, as computed from some available declination measurements, can be made.

The Central Mojave domain. The late Neogene crustal deformation in the Mojave Desert is predominantly accommodated by sets of NW-trending right-lateral strike-slip faults and E-W-trending left-lateral strike-slip faults (fig. 7) (Dibblee, 1961; Garfunkel, 1974; Dokka, 1983). On the basis of structural data, Garfunkel (1974) suggested that deformation in the Mojave is accommodated by contemporaneous right-lateral strike-slip faulting and counterclockwise (CCW) block rotation.

The anticipated rotation in the Central Mojave domain is counterclockwise. The magnitude
of block rotation is estimated to be about 30" (CCW) using GarfunkePs (1974) fault slip estimates and around 15" (CCW) using Dokka's (1983) more conservative fault slip values. Preliminary paleomagnetic data from this region (unpublished, Morton and Hillhouse, 1983) suggest about 15" CCW rotation of the Mojave blocks during the last 6 Ma. Paleomagnetic data from nearby southwestern Arizona (Calderone and Butler, 1984) show CCW rotational 13.7" .

Seismic data, active surface deformation, and geodetic data, may also be consistent with crustal deformation being accommodated by simultaneous strike-slip and rigid block rotation even at present. In 1975, the Galway Lake earthquake (M = 5.2) (fig. 9) was accompanied by a 6.8 km long surface rupture with right-lateral displacement (Hill and Beeby, 1977) on a previously unmapped fault trending N20°W. Remarkably this displacement did not occur along the established nearby Emerson Fault, which trends N 45" W. Sauber et al. (1986) concluded that the preferred nodal planes for five out of nine recent earthquakes in this area are also oriented 20" to 35" away from the general strike of the Mojave faults, roughly in the direction of the Galway rupture direction, although they consider this direction of faulting to be "secondary." Furthermore, the fault plane solutions for the larger shocks of the 1979 Homestead Valley earthquake sequence (Hutton et al., 1980) show nodal planes trending in N 4" W, again 30" to 40" away from the strike of the older Mojave fault trend (Stein and Lisowski, 1983). It is remarkable that both ruptures tend together to delineate a single line which combined suggest a

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single fault at least xx km long.

The direction of faulting in the central Mojave domain of the Galway Lake and Homestead Valley earthquakes might very well be part of the development of a new fault system, which is gradually replacing the older, now rotated out of favor strike-slip faults here. This interpretation is remarkably consistent with the direction of the compressive stress in this area which is horizontal, at approximately AT 30° E to AT45° E as given by Zoback and Zoback (1980) and as inferred most recently from breakout measurements in the Cajon Pass well (M. Zoback, personal communication). The older Mojave faults are quite clearly unfavorably oriented relative to this stress direction, whereas the Galway Lake-Homestead Valley direction is quite favorable, mechanically speaking.

The Lake Mead strike-slip fault systems. The Lake Mead fault system (LMFS) consists of a few major NE-trending left-lateral strike-slip faults (fig. 10). The accumulated offset across these faults is about 65 km, as obtained from offset of Late Neogene volcanic rocks (Anderson, 1973; Bohannon, 1979). Based on the ages of the displaced rocks of the Hamblin-Cleopatra volcano (Anderson et al., 1972) strike-slip faulting here post-dates 11 Ma.

A detailed study of the geometry and nature of faulting in the LMFS was carried out by Angelier et al. (1985), in the Hoover Dam area. Their results highlight the presence of a second fault system trending roughly NW, divided into two distinct groups: right-lateral strike-slip and normal dip-slip. Of these it appears that the NW-trending right-lateral faults have also participated in the deformation of the region. These faults, typically 5 to 20 km long, are roughly conjugate to the major NE left-lateral faults, which are typically 50 to 150 km long.

Paleomagnetic data from the Hamblin-Cleopatra volcano were obtained (Ron et al., 1986) from 22 cooling units of volcanic rocks with K-Ar ages by Anderson et al. (1972) of 11.3 Ma and 12.7 Ma. The outward radial distribution of the lava flows attitude were used to control tectonic tilt. The paleomagnetic results reveal little if any rotations about the horizontal axis, corresponding to the lack of any vertical rotations associated with normal fault slip. In contrast, the declinations of the in situ field directions yielded large horizontal counterclockwise rotation of -29.4 ± 8.5 about the vertical axis. This counterclockwise sense of rotation is surprisingly consistent according to the block rotation model with the left-lateral slip on the NE-trending major fault set. In contrast, this sense of rotation is to be expected if the associated slip took place on the shorter pervasive NW-trending right-lateral faults of the LMSZ. Our interpretation of the results is shown in figure 11, in which the major NE fault coincides with the fixed boundary direction of this domain which does not rotate with rotation totally associated with the conjugate NW faults.

Supporting evidence for this interpretation comes from the active faulting in the LMFS, as indicated by regional seismicity. According to Rogers and Lee (1976) the majority of epicenters recorded are associated neither with the smaller NW-trending faults nor with the larger NE ones, but rather with younger and shorter, N-S-trending fault segments, (fig. 10). Focal mechanisms for these earthquakes reveal right-lateral strike slip motion on these faults (Rogers and Lee, 1976).

Figure 11 presents our current model for the fault geometry, the sens of horizontal slip, and
the nature of block rotation in the LMSZ, beginning about 11 m.y. ago and still in progress now. Left-lateral shear in Miocene times caused local right-lateral horizontal displacements on a local set of faults initially trending N-S. These right-lateral displacements lead to an approximately 30° counterclockwise rotation of the blocks and their bounding faults resulting in their present NW direction in which became locked (Nur et al., 1986). Assuming that the stress orientation remained constant during the past 11 m.y. (Zoback and Zoback, 1980; Carr, 1984), subsequent deformation by right-lateral shear was accommodated by a new set of more favorably oriented N-S faults which are active today.
This interpretation provides a link between the structural, paleomagnetic and seismic data and integrates them into a single continuous process of older and current crustal deformation by strike-slip and block rotation. Although the results are specific to the LMSZ, the model by

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analogy should be equally applicable to other Nevada strike-slip fault systems. Because this generalization could have profound impact on our understanding of the neo- and active tectonics of important portions of the Great Basin, much more work is needed in the future.

(c) The San Gabriel Mountain domain. The formation of multiple strike-slip fault sets, associated with clockwise rotation can explain some of the complexity of faulting in the San Gabriel Mountain domain (fig. 12). Dibblee (1961) has long ago recognized that the set of subparallel right-handed strike-slip faults here imply clockwise rotation of both blocks and faults. Subsequent paleomagnetic results by Terres and Luyendyk (1985) revealed such rotation, with magnitudes reaching 50° to 55° . As suggested on the basis of the mechanical analysis of Nur et al. (1986) that such rotations may be larger than could be accommodated by a single fault set. As shown in figure 12, after Carter (1982), there is good evidence for the presence here of an older strike-slip fault set, which is cut and offset by the younger set. According to the block rotation mechanics model, this older set, oriented now roughly NE, has rotated gradually clockwise, until it became locked due to the increasing normal stress and decreasing shear stress, with further deformation being accommodated by the new set of faults. The observed angles between the two sets is in the range of 30° to 40" , in good agreement with the predicted value of Nur et al. (1986).

4. RELEVANCE TO GEODYNAMICS

Scale dependence and mechanics. The block rotation concept may be important for the interpretation of crustal deformation data e.g. in the Western United States, such as obtained from repeated geodetic measurements. The process of block rotation implies, for example, the strain, fault slip, and crustal deformation must generally be scale-dependent. On the scale of a few kilometers, deformation is controlled by fault spacing; on the scale of a few tens of kilometers, deformation is controlled by fault domains; and on a scale of a few hundreds of kilometers, deformation is controlled by regional tectonics and plate motions. For example, a local domain can have left-lateral slip and clockwise rotation in an overall region with right-lateral and hence counterclockwise rotations. Also the presence of multiple sets implies that faults are generally not in "optimal" directions, and may thus be ambiguous regional stress and deformation indicators.

In order to quantitavely incorporate the roles of fault sets and domains in crustal deformation and evaluate their importance, their mechanics must be understood in some detail. For example rock strength and frictional properties in two and three dimensions must be considered in the formation of fault sets, domains, and domain boundaries. These properties may also control the formation of multiple fault sets of a given type (strike-slip, normal, thrust) or mixed types (strike-slip and thrust in the Transverse Ranges, strike-slip and normal in the Basin and Range). Consequently slip on a single fault, say during a single earthquake, such as the Galway Lake or the Homestead Valley events, does not necessarily mimic the regional deformation. Similarly, the deformation of a given domain does not mimic regional deformation: For example, the overall deformation of the Mojave region is the sum of the deformations of its three main domains. Clearly the strain of the central domain must differ markedly from the strain of the NE domain.

Deformation of a single domain. As slip on and rotation of the faults of a single domain
set progress with time, the local shear stress required to cause distortional strain of the domain must increase, until, as shown earlier, it becomes more favorable to create a new set of faults. In the case of two-dimensional strike-slip faulting preliminary analysis yields the following relation between the shear stress T required to cause fault slip, and block rotation

r=r0 ^ (5)

where r0 is the shear stress for the initial, mechanically most favorable orientation of the fault set,

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and <f> (0 < <^» < <j> e) is the magnitude of block rotation.

The generalized extensional strain e is given by

sna -

where / is the domain length, IQ is the initial length (when rotation is <^» = 0), and a is the initial orientation of the faults relative to the direction of shortening. This stress-strain relation for a single domain can be used to relate local deformation to tectonic stress.

Crustal deformation of a cluster of domains. In figures 13 we show cartoonwise one of
the problems which can arise in the interpretation of horizontal crustal deformation data when strike-slip-type block rotation is involved, especially when measurements of azimuth changes are concerned. Consider two geodetic stations, A and B, first within a single domain (fig. 13a). As the domain deforms (fig. 13a) each station undergoes rotation of the same sense. This case is essentially identical to continuous deformation, and no special problems arise. However if station A is in one domain, and station B in a second (fig. 13b), a major difference arises from the homogeneous case, since the line from A to B, in our example, will rotate counterclockwise, whereas the line from B to A will rotate clockwise. If the fact that the geodetic array spans a domain boundary is not inlcuded in the analysis and the interpretation of the geodetic data, large and fundamental errors can be introduced. In the example here the two rotations would tend to cancel one another, leading to totally erroneous results, regarding shear strain.

5. RELEVANCE TO EARTHQUAKE PREDICTION

Although the concept of block rotations is directly applicable to the understanding of distributed faulting, it is not immediately obvious whether and how it might be related to earthquake prediction. However two applications are worth exploring: the initiation and arrest of earthquake rupture, and the prediction of which fault will slip.

As suggested by a number of investigators, irregularities in the geometry of a fault or fault system may be" the cause for either or both the initiation of earthquake ruptures and their arrest. Some models envision that rupture begins at a point of low strength or resistance to slip and is stopped at strong points, or barriers. Other models envision that rupture actually begins at strong points, and spreads into weak, or compliant regions where it is arrested. Regardless of model, it is apparent that large heterogeneities of fault strength or normal and shear stresses are required, and that such heterogeneities may be related to geometrical irregularities along faults and fault systems.

As pointed out by e.g. Freund and Merzer (1976) and Nicholson et al. (1986), the rotation
of block within fault systems, and/or the rotation of the fault strands themselves may be attractive candidates for the origins of these mechanical variability along fault. The study of the details of domains of block rotations and their development in time and space may consequently be of significant for the understanding of both rupture initiation and stopping, and thus the prediction of where large earthquakes may occur.

The concept of blocks and their role may also be important in predicting the particular fault
on which a future earthquake will occur. This question is of special interest because the most successful earthquake prediction to date the Haicheng, China, earthquake of February 5, 1975 involved the correct prediction of the time and the approximate of the event, it failed to predict the actual fault on which this earthquake occurred. The epicenter instead of falling on the major through-going NW right-handed fault, actually fell on a conjugate(?) left slipping fault, or a fault in a conjugate set or domain. It may well be that the stress required to rupture this fault was unusually high, due to the geometry of the situation, so that the associated precursors were unusually strong.
A second example is the M ~ 5.0 Mount Lewis earthquake of 1985 east of the San Francisco Bay, California. Although the main strands of the San Andreas here, with right-handed slip,

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trend NW-SE, many active faults are present, possibly due to the complex of pull-aparts and pushups (Aydin and Page, 1984; Aydin and Nur, 1982) which are pervasive here. This earthquake fault plane solution indicates handed slip on an anomalous N-S- not NW-SE-trending fault. As shown in figure 2, the block rotation model may provide a simple explanation for this otherwise puzzling type of slip. According to the model, this earthquake may be associated with a particular fault out of a set of N-S-trending strike-slip faults in an east bay domain bounded by the two major San Andreas strands, the Calaveras and Hayward faults.

6. CONCLUSIONS

The kinematics of block rotations bound by fault sets in domains, as first proposed by Freund (1970) provide an attractive model for understanding and anlyzing crustal deformation when pervasive faulting is involved. The model is especially important because crustal deformation by pervasive faulting may very well be the rule.

Mechanical considerations of rock and fault strength and fractional properties suggest that new fault sets may form when older ones have rotated sufficiently away from the direction of optimal failure. Thus the combination of concept of block and fault rotation with fault mechanics provides a simple model which can explain rather complex fault patterns in situ.

The rotation of blocks and faults due to fault slip in a distributed fault systems must be included in the analysis of geodetic and other crustal deformation data, because it can modify the interpretation of such data significantly. Finally, the mechanics of block and fault rotations may be intrinsically linked with those heterogeneities of fault systems which are responsible for the initiation and arrest of rupture. This is true especially in distributed fault system, where the prediction of an earthquake involves not only time and magnitude, but also the actual fault on which the event is to occur.

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Figure Captions

Figure 1. Mojave domains.

Figure Ib. Details of the central Mojave Desert (CMD) and eastern Mojave Desert (EMD). Each domain consists of roughly parallel fault sets of consistent sense of slip, which is expected
to be directly associated with block and fault rotations: counterclockwise rotation with right-lateral slip in the CMD, and clockwise rotation with left-lateral slip in the EMD.

Figure 2. A 2-D model which illustrates the simultaneous activity of strike-slip displacement and rotation of the faulted blocks, (a) The initial configuration; (b) after deformation. A set
of left-lateral faults rotates clockwise and a set of right-lateral faults rotates counterclockwise.

Figure 3. Fault set kinematics. Geometrical relation between spacing between faults u>, their initial orientation a, initial length / 0 to slip d and rotation <£.

Table 1. Determined rotations.

Figure 4. Mohr circle representation of state of stress in rock with existing fracture: 0 = tan"1 /* is angle of friction, 50 and 5! are cohesive strength of virgin rock and fracture, respectively, and <f> e is largest permissible angle between existing fracture and direction
in which new shear fracture will develop if further rotation takes place.

Figure 5. Maximum permissible angle <j> e between sets or rotating shear faults as function of cohesive strength of rock S0 and fault S l} and effective overburden pressure a0 for /* =
0.6 and n = 0.

Figure 6. Modified block model, showing new fault set required to accommodate block rotation greater than 45" .

Figure 7. Fault domains in northern Israel, showing fault sets, directions of slip, and paleomagnetically determined rotations in these domains.

Figure 8. Comparison between structurally derived and paleomagnetically determined rotations in northern Israel (see fig. 7). Solid line represents perfect agreement; dashed line is
linear best fit to data.

Table 2. Angles between sets.

Figure 9. The formation of a new fault set. The faulting directions associated with the 1975 Galway Lake earthquake and the 1979 Homestead Valley earthquake sequence. Note
their azimuths which are distinctly different from the older, well-established central Mojave faults, and their alignment along a single trend.

Figure 10. Map of study area, showing Hamblin-Cleopatra volcano displaced laterally by Hamblin sinistral fault, system of coaxial radial dikes, attitude of primary dips on lava
flows (after Anderson, 1973), and locations of paleomagnetic sampling sites.

Figure 11. Simplified geometric model for deformation of left-lateral Lake Mead fault system by right-lateral strike-slip faults and counterclockwise rotation. Direction of elongation is west-northwest, which is close to direction of minimum horizontal stress of Zoback and Zoback (1980). (a) Stage I. Slip on the major NE-trending faults; (b) Stage II. Slip and rotation on the shorter NW-trending fault; the original orientation of these faults, based

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on the measured rotation was N-S; (c) Stage III. Slip and seismicity on the active N-S-oriented faults. The slip was presumably transferred to these faults after the NW-trending set became locked due to their rotation.

Figure 12. Multiple strike-slip fault sets in the San Gabriel Mountains (after Carter, 1982). Note the younger strike-slip NE-trending faults offsetting the older E-W-trending strike slip.
Both sets have the same left-handed sense of motion and are therefore not conjugate
sets. The paleomagnetically determined clockwise rotation of 53" (Terres and Luyendyk, 1985) is consistent with the observed left-handed slip.

Figure 13. Sketch illustrating the influence of block rotation and domains on azimuthal changes, as determined from geodetic measurements. Heavy arrows indicate original lines connecting stations A and B. Light arrows indicate sense of fault slip.

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REPORTED BLOCK ROTATIONS INFERRED FROM PALEOMAGNETIC AND STRUCTURE DATA

Location of
rotated blocks
Iran-Sistan
Jordan Valley-Israel
Mojave-California
Mojave -California
Southwest Arizona
(Mojave blocks)
Imperial Valley, California
Morro Rocks-lslay Mill
complex, California
Eastern Klamath belt,
northern California
Southwest Washington
Northern Israel

Hermon and Lebanon,
northern Israel and
Lebanon
Rio Grande rift,
New Mexico
Dixie Valley,
west-central Nevada
Lake Mead, Nevada
San Gabriel region,
California
Channel Islands,
California
Santa Ynez Range.

Magnitude and
sense of rotation

± 45«(S)
-25, -72« (P)
30° (S)
-16« (P)
14' (P)

+30-+700 (Sd)
+49° (P)

483-+116* (P)

420-+650 (S,P)
423-+360 (S,P)
24-53°

59° (P,S)

+12, -16° (P)

36° (P)

27° (P)
+53° (P)

+53° (P)

+36-+920 (P)

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Garfunkel, 1974
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Fagin and Gose, 1983

Wells and Coe, 1985
Ron et al., 1984

Ron and Nur, 1985

Brown and Golombek, 1985

Hudson and Geissman, 1985

Ron et al., 1986
Terres and Luyendyk, 1985

Kamerling and Luyendyk, 1985

Hornafius, 1985
Note: P«paleomagnetic; S -structure; Sd- surf ace deformation; counter-clockwise; + clockwise.

Table 1.

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~^ !

Figure 1. Details of the central Mojave Desert (CMD) and eastern Mojave Desert (EMD). Each domain consists of roughly parallel fault sets of consistent sense of slip, which is expected
to be directly associated with block and fault rotations: counterclockwise rotation with right-lateral slip in the CMD, and clockwise rotation with left-lateral slip in the EMD.

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DOTATION OF BLOCKS BORDERED §Y TWO SETS Of STRIKE SLIP FAULTS

A. Initiol configuration
6. Afltr deformolion

B

Figure 2. A 2-D model which illustrates the simultaneous activity of strike-slip displacement and rotation of the faulted blocks, (a) The initial configuration; (b) after deformation. A set of left-lateral faults rotates clockwise and a set of right-lateral faults rotates counterclockwise.

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Shear
Stress

T Virgin Rock

Existing
fracture

00

3 ^<Pc V0 a1

Normal stress cr

Figure 4. Mohr circle representation of state of stress in rock with existing fracture: 0 = tan"1 /* is angle of friction, S 0 and Si are cohesive strength of virgin rock and fracture, respectively, and <j> e is largest permissible angle between existing fracture and direction
in which new shear fracture will develop if further rotation takes place.

Initial
Configuration

00
h-»

ON

B

Incipient fault set

Incipient
fault

Locked first
set

Active
second
set

Figure 6. Modified block model, showing new fault set required to accommodate block rotation greater than 45" ,

35* 35° 30'

33*

32
30'

SOUR

33*

32°
30'

Figure 7. Fault domains in northern Israel, showing fault sets, directions of slip, and paleomagnetically determined rotations in these domains.

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K->

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Clockwise Rotation

(ten lateral faults)

(STRUCTURAL)

40°

20°-

Anticlockwise Rotation

(Right lateral faults)

20° 40°

(PALEOMAG.)
- » I
60<

Figure 8. Comparison between structurally derived and paleomagnetically determined rotations
in northern Israel (see fig. 7). Solid line represents perfect agreement; dashed line is
linear best fit to data.

Figure 9. The formation of a new fault set. The faulting directions associated with the 1975 Galway Lake earthquake and the 1979 Homestead Valley earthquake sequence. Note
their azimuths which are distinctly different from the older, well-established central Mojave faults, and their alignment along a single trend.

|TjTv|Td7,l Tertiary rocks of Hamblin-Cleopatra Volcano Ti intrusive
TV volcanic
Td dike
±; fault

30 strike and dip of lava flows

Paleomagnetic sampling sites

Figure 10. Map of study area, showing Hamblin-Cleopatra volcano displaced laterally by Hamblin sinistral fault, system of coaxial radial dikes, attitude of primary dips on lava flows (after Anderson, 1973), and locations of paleomagnetic sampling sites.

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on the measured rotation was N-S; (c) Stage ffi. Slip and seismicity on the active N-S-oriented faults. The slip was presumably transferred to these faults after the NW-trending set became locked due to their rotation.

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MULTIPLE STRIKE SLIP FAULT SETS

IN THE SAN GABRIEL MOUNTAINS SOUTHERN CALIFORNIA

(After Carter: 1982)

0 1 2 3 4 5 Km

1 Pole Canyon fault
2 Oak Springs fault
3 Magic Mountain fault
4 Transmission Line fault

5 Mt. Gleason fault
6 fox Creek fault
7 Mill Creek fault

A Soledad fault
B Lonetree fault
C Slaughter Canyon fault

Older faults
Younger faults

Figure 12. Multiple strike-slip fault sets in the San Gabriel Mountains (after Carter, 1982). Note
the younger strike-slip NE-trending faults offsetting the older E-W-trending strike slip.
Both sets have the same left-handed sense of motion and are therefore not conjugate
sets. The paleomagnetically determined clockwise rotation of 53" (Terres and

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Figure 13. Sketch illustrating the influence of block rotation and domains on azimuthal changes, as determined from geodetic measurements. Heavy arrows indicate original lines connecting stations A and B. Light arrows indicate sense of fault slip.

THE MECHANICS OF THE INITIATION REGION OF THE 1984
MORGAN HILL, CALIFORNIA, EARTHQUAKE.

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Abstract

The seismic behaviour and spatial distribution of slip in
the region along the Calaveras fault where the 1984 Morgan Hill, California, earthquake initiated is quantified by converting magnitude estimates to seismic moment and then to spatially-distributed slip. It is found that the slip contribution from shocks in size i^anges separated by factors of two in mean fault dimension varies by factors of two. These observations, and the spatial distribution of slip, are in good agreement with the predictions of fractal theory for pure brittle failure of a fragmenting bend. In contrast, the seismic behaviour and slip distribution of a. well-established nearby creeping region in the Hollister Valley are not consistent with these theories, except for very small earthquakes. A striking observation about the Morgan Hill earthquake is that it initiated in a small local slip minimum but then propagated and released most of its energy in a larger minimum nearly ten kilometers to the south.

Introduction

The problem of predicting earthquakes can be approached from
a number of perspectives ranging from the probablistic correlation of varied phenomena to model dependent descriptions using limited or selected data sets. We are not attempting here to consider earthquake prediction in these ways but instead to discuss aspects of the mechanics of the earthquake process. However, clarification of the mechanics of earthquake initiation presumably will assist all approaches to the prediction problem. In an earlier paper Bakun, King and Cockerham (1986) examined the behaviour of the Calaveras fault zone by comparing the slip that has been associated with earthquakes of all sizes with the motion determined from geodetic measurements. We now look in detail at the region where the M = 6.2 Morgan Hill earthquake of April 24, 1984 initiated. To highlight certain

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features we apply the same techniques to a part of the Hollister Valley. The iMorgan Hill earthquake initiation region is associated with a fault bend so that the brittle fragmentation processes at tectonic knots or bends, discussed by Sykes and Nisheriko (1984), Scholz (1985), Nishenko and Williams (1985) or King and Nabelek (1985), might be verified. The Hollister Valley region, on the other hand, is known to fail primarily by fault creep so that relaxation processes might be expected to dominate (Husseini et al. 1975, King 1986).

38°00

36°

Morgan Hill
zone '

~\ Coyote \ Lake 0" "

I zone - .

MflGNITUDE SYMBOL
5.0 TO 6.0- O
4.0 TO 5.0- o
3.0 TO 4.0- o
2.0 TO 3.0-
LESS THftN 2.0

50 KILOMETERS

123°00 121°00 onn'

Figure 1. Seismicity (magnitude > 1.3) in the San Francisco Bay area for 1976-1984. The solid polygons enclose the region of epicenters of earthquakes considered in this study. Distances are measured in kilometers from A to A'.

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a) MAP CALAVERAS FAULT 1969 - 1984
CONCORD
FAULT

1979 COYOTE LAKE
RUPTURE ZONE

1984 MORGAN HILL
RUPTURE ZONE

20 40 60 80 100 DISTANCE (KM)

120 140 150

Figure 2. Seismicity within the areas indicated in figure 1 for the perid 1969-1984. (a) map view and (b) vertical cross-section Solid enclosures indicate the region of this study.

King (1983) suggested that the logarithmic relations between earthquakes of different sizes are a consequence of 'fragmentation' produced by continuing motion at a fault bend in brittle material. This leads to a predicted b-value for the magnitude frequency relation of 1, the globally observed value. The argument that leads to this conclusion adopts only geometric principles (Mandelbrot 1977, King 1983). In addition to predicting the observed b-value relationship, the arguments suggest that certain spatial relations should exist between earthquakes of different sizes. This study investigates such spatial behavior.

Data analysis

The data used in this study are a subset of the data used
by Bakun et al.(1986). All the epicenters for the San Francisco Bay area are shown in figure 1 with the two regions that form the subject of this study outlined with solid lines. Figure 2 shows the same region and also a cross-section onto which all the shocks are projected. Bakun et al. (1986) discuss location accuracies, catalog completeness, the reliability of magnitude estimates, and the procedures to convert the earthquake catalog to slip distributed on the fault.

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a

DISTANCE(km)
80 90
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CNSTANCE(km)

towest contour" for next higher
magnitude range

"Lowest contour" in frame

"Lowest contour" x2

lowest contour" x4

Figure 3. Summed slip in different magnitude ranges for the Morgan Hill earthquake initiation region:

20 Moment {xlO dyne-cm.) Lowest contour (cm)
0.156- 1.25 0.02
1.25 - 9.99 0.04
9.99 - 79.90 0.08
79.90 - 641.0 0. 16
641.0 -5310.0 0.32

(f) Slip for the Morgan Hill earthquake derived in the mariner discussed by Bakun et al. (1986). The contour intervals are factors of two except for (f) which has intervals with factors of four.

Magnitude
(a) 1.83-2.58,
(b) 2.58-3.33
(c) 3.33-3.94
(d) 3.94-4.54
(e) 4,54-5.14

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Bakun et al. (1986) summed the slip for events of all sizes
for different time periods. Here we look at the complete 1969- 1984 time period, but examine the slip distribution as a function of earthquake size. We consider earthquakes in bands of moment, that correspond to earthquake faulting that varies in

Summed slip
(cm)

100.0-r

10.0-

1.0-

0.1-

0.01-

-(0.01)x212

(0.

(0.01)2f
0.42

0.22

0.11

(0.

0.045

0.021

1234567

Magnitude

Figure 4. Summed slip in the magnitude bands of figure 3 in the depth ranges 5-12 kilometers. The magnitude bands are indicated by bars with the amount of slip in cm shown above each bar. The vertical axis is logarithmic slip in base 10 and base 2. The open circle is the Morgan hill slip. See the text for the significance of the solid circle.

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characteristic dimensions by factors of two. If slip is proportional to fault dimension, then the corresponding seismic moment bands are separated by factors of eight. We plot slip logarithmically in figure 3 with each contour representing an increase of slip by a factor of two. In each case the first, contour plotted, the "lowest" contour, is a factor of two in slip greater than the "lowest" contour in the preceeding frame (i.e. the adjacent lower moment band). Note that the shaded regions in figure 3 have similar areas so that, on average, the slip for successive frames increases by factors of two.
The slip shown in figure 3f is for the Morgan Hill main shock. While the lowest slip contour is double that of figure 3e, the contour interval changes by factors of four and not two. The total slip in figure 3f is consequently much greater than twice that of figure 3e. The Morgan Hill slip distribution shown follows Bakun et al. (1986).
On each figure, the outline of the lowest slip contour for
the following figure is outlined by a dotted line to assist in comparing the relative location of slip for each frame. The figures show that peaks in adjacent moment bands rarely coincide. However, the slip in adjacent moment bands is not unrelated. Slip in a moment band occurs in the troughs or on the flanks of slip at the next higher band.
To quantify the slip in the different size ranges we summed slip in the range of depths between 5 and 12 kilometers, the depth range within which Bakun et al. (1986) found most seismic slip. The total slip for the different size bands is shown in figure 4. The horizontal bars show the magnitude ranges corresponding to the fault length bands. The vertical axis is slip on a logarithmic scale labeled both in base 10 and in binary units. The former is chosen so that absolute values can be read. The latter is added to show the approximate doubling of slip for each increase of mean fault length by a factor of two.
Caution is necessary to assess the significance of the slip
for the Morgan Hill earthquake shown in figure 4 by an open circle. Since the earthquake slip is not well defined spatially and since it extends over an area larger than our study area, the slip value obtained is not very accurate. We estimate that the apparent slip could be reduced by perhaps a factor of two if the area of study were chosen differently. Nonetheless, the Morgan Hill slip would still be more than a factor of ten greater than would be expected from the trend of smaller events. One explanation for this discrepancy is that the summed smaller event slip represents only a small part of the seismic cycle, so that

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D'S'ANCE(km)
90
l I I I I I I I I

100

5-

E ~

10-
Q_
LJ

Q

15-

Figure 5. (a) The summed slip in the initiation region of the Morgan Hill earthquake (i.e. the first five frames of figure 3) for all magnitudes but excluding the Morgan Hill main event slip, (b) The slip of the Morgan Hill earthquake superimposed on the upper figure. The star is the Morgan Hill earthquake hypocenter.

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including the total Morgan Hill slip is unreasonable. Bakun et al.(1986) argue that the return period for the M=6 Coyote Lake and Morgan Hill shocks is about 80 years whereas the data in this study comes from a 16 year time window (1979-1984). Hence it is reasonable to scale the Morgan Hill slip by reducing its slip in proportion to these time intervals. This "time-scaled" slip is shown in figure 4 by a solid circle. Even with this adjustment, and even allowing for the effects outlined earlier, the Morgan Hill slip lies a factor of two or more above the trend. This observation, equivalent to noting that the largest event on a fault apparently has a magnitude significantly above the b-value trend, has been documented elsewhere (e.g. Wesnousky et al. 1985; Singh et al. 1984.)
Of obvious interest is the relation of the main event to the
sum of all the lower magnitude seismicity. This is illustrated in figure 5a where contour intervals are in factors of four and the lowest contour is a slip of 0.04 cm. Figure 5b shows that the Morgan Hill hypocenter (indicated by a star) lies in a minimum in the earlier seismicity, and rupture in the main event propagated towards, and filled, a much more extended slip minimum to the southeast.
The same processes that we have applied to the data from the Morgan Hill initiation region were also applied to the seismicity in the Hollister Valley region. Plots equivalent to those in figure 3 are shown in figure 6. The Hollister Valley figures are markedly different in character from those for the Morgan Hill region. First, the seismicity in different size windows is generally unrelated in space; only a few examples of events occurring on the flanks or minima of adjacent scale seismicity can be seen and only at small magnitudes. Second, the slip in different scale sizes does not bear the simple binary amplitude relations seen for Morgan Hill; one intermediate scale range (figure 5d) is completely devoid of events. The slip amplitude versus magnitude window plot in figure 7 emphasizes this observation.

Discussion

A body subject to displacement boundary conditions that conserve volume must permit simultaneous shear motion on a number of planes (five or more) with different orientations. Consequently a deforming rock mass, whether on the scale of a

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Summed slip
(cm)

0.1

(0.001)x2'

0.01

°£Lio.oi2

<0.001)x23 0 0056
0.0049.

0.001 o.oo
5 6

Magnitude

Figure 7. Summed slip in the magnitude bands of figure 6 in the depth range 5-12 kilometers. The magnitude bands are indicated by bars with the amount of slip in cm showm above each bar. The vertical axis is logarithmic in base 10 and base 2.

rock sample or a continent, must contain faults of different orientations. This is a demand of geometry and not of rock properties. The problem has been discussed by King (1983), Nabelek and King (1985) and Turcotte (1986) who suggest that faults arrange themselves in a binary scale hierarchy. To conserve slip at a bend, for each drop in scale size by a factor of two, the slip that must be accommodated by smaller faults must drop by a factor of two. When moment is converted to magnitude, the relation becomes identical to the Gutenburg-Richter relation with a b-value of unity.

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The preceding derivations depend on the properties of a three dimensional system of faults. The complete system is difficult to represent pictorially. Analogous two dimensional systems can readily be drawn (e.g. King 1983) but they do not have all of the geometric properties of a three dimensional system. Alternatively, a fractal algorithm can be described that will produce the three dimensional system but this is not very helpful visually. Figure 8 is a sketch that has some of the properties of a three dimensional system. We can interpret figure 3 as showing a series of sections through a system of faults with a geometry similar to that shown in figure 7. The small events fall at the edges of the larger events and the relative proportion of events of different sizes is close to that predicted. The rupture of the large event (figure 5b) initiates from within the network of small faults at the bend.

Figure 8. A schematic representation of some of the faults in a fault bend region. Note that only three of the predicted five fault orientations are shown. Because the bend is small the total slip and moment release on these planes will be small. Nonetheless they are of mechanical significance.

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The data that we present is compatable with a "fragmentation" model but a result from Bakun et al. (1986) possibly makes this conclusion surprising. We found, when comparing motion determined from geodetic data with that determined seismically for the same region, that at most 25 percent and possibly as little as 5 percent of the deformation is associated with seismic radiation. We concluded that the remainder must be accommodated either as aseismic fault slip or as volume creep in the body of the surrounding rock. There is no obvious reason why creep processes should give rise to the same fractal distribution as that apparent for earthquakes.
However, it is apparent that the deformation process does
not disturb the geometrical relations between events of different sizes in the Morgan Hill earthquake initiation region. The results we have obtained in a known creeping region are very different (figures 6 and 7). Except for the smallest events, no clear evidence for a "fragmentation" style of deformation exists. We do not at present know how to offer testable theories to explain these differences.

Conclusions

The bend region where the Morgan Hill earthquake initiated behaves in a manner that is consistent with the predictions of a simple fractal "fragmentation" model for brittle material. However, an earlier study appears to suggest that substantial creep occurs in the same region. We do note that the Morgan Hill earthquake behavior is substantially different from a region of known fault creep. We are challenged to consider ways in which the relations between creep and seismic slip in such regions can be examined more closely.
A particularly interesting result of our earlier study which
our new figures clarify is that the Morgan Hill earthquake initiated in a local slip minimum but released most of its moment when rupture propagated to a much larger slip deficit to the south-east (figure 5). The initiation point and the main energy release apparently occurred in separate seismic gaps separated by nearly 10 kilometers.

Acknowledgments

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On Noise Levels in Deformation Measurements: Comparison in the Frequency Domain

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Abstract

To evaluate the usefulness of different measurements for earthquake prediction, we must consider not just their sensitivity to precursors but also their noise level. Because the physics of possible precursors are poorly understood, often we can quantify only the noise. For many methods, notably instruments such as strainmeters, the noise level varies as a function of frequency, and is best described using spectral analysis. Measurement error from geodetic surveys may be cast into this form if the time between surveys is con-sidered. Spectral comparisons show over what time scales different measurements have the least noise; with existing techniques, the crossover time between geodetic techniques and observatory-based instruments is from months to years.

1. Introduction

One of the fond hopes of earthquake prediction research has been that since earth-quakes are caused by the release of stored strain energy, direct measurements of deforma-tion might yield clear precursory signals. In pursuit of such precursors, there have been several large-scale programs (most notably in Japan) to measure strain and tilt changes. While it must be admitted that no clear successes have emerged from these efforts, it may be argued that this is caused not by the absence of precursors but by the signals being small relative to the noise of the instruments used. At present, this noise is high enough that strain and tilt measurements have apparently not recorded any part of the Earth's deformation other than coseismic offsets and seismic waves, secular accumulation, and (perhaps) postseismic adjustment. Until noise levels are lower we cannot reject the possi-bility of strain or tilt precursors to earthquakes.
A consequence of the above reasoning is that a major focus in this area of earth-quake prediction should be on the noise levels attainable with different measurement tech-niques. (Note that this situation is quite different from seismic precursors, in which we have no trouble measuring our basic data, namely earthquake occurrence.) I attempt to discuss this here, specifically by developing a method for comparing noise levels from apparently very different types of measurement. The emphasis is on the temporal behavior of the noise, since this is what matters for precursor detection.

2. Data Types
Without undue simplification, we may separate deformation measurements into two groups. Before distinguishing them, we note that they have one thing in common: they consist of samples at some roughly constant interval Af, and therefore cannot provide meaningful information about behavior at frequencies above the Nyquist frequency

fn - %M - (fj- 1

where tn is a defined "Nyquist period". This is an unavoidable property of any actual set of time-series data.

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Because we are concerned with noise, our separation of measurement types will be according to the error model that is appropriate for them. The basic distinction is whether the errors associated with each output value are independent from one value to the next or correlated between values. As we will see in Section 5, the notion of indepen-dent errors is an idealization; but it is a sufficiently good one to make it worth pursuing.

By design, this distinction separates two familiar classes of deformation measure-ments. It is generally assumed that geodetic measurements have independent errors. These techniques provide an absolute measure of some part of the difference in location between two stations. They include electronic distance measuring (EDM) systems, giving the distance between two points; "space techniques" (such as VLBI, SLR, or GPS), which involve measurements from both sites to a common source and give the vector difference between locations in all three coordinates; and spirit leveling, which gives the difference in geopotential elevation between two points. The measurements that clearly possess corre-lated errors (e.g., drift) are those from higher-sensitivity instruments. Often these are called "observatory-based"; they include all known types of strainmeters and tiltmeters, along with proxy indicators such as water level.

We shall assume for both sorts of measurements that the errors are stationary; this gives a considerable gain in simplicity without undue loss of generality. It also has an important corollary. For independent stationary errors the error distribution does not change with time: this means that, in the absence of deformation, repeated measurements will not show any long-term divergence from their initial value. On the other hand, if the errors are correlated the measurements may diverge from their initial value; while this cannot go on forever without departing from stationarity, it may continue over any finite interval (Keshner, 1982).

The absence of long-term divergence is an implied property of absolute measure-ments, which have a known zero point. The geodetic methods referred to above are all absolute techniques. Usually the price of higher sensitivity, as in the observatory-based instruments, is that the measurement is a relative one. It should be noted that even an absolute instrument (such as a water-level tiltmeter) can be subject to long-term diver-gence. I shall, however, use the terms "absolute" and "relative" below for the two types of measurements, since "independent" and "correlated" are needed for statistical descrip-tion.

Clearly, deformation measurements differ in other ways than just the characteristics specified above. Most notably, geodetic measurements, being over longer distances, pro-vide a spatial coverage and freedom from local distortions that an observatory, or even an array of observatories, cannot match. While such spatial coverage is important for many studies of crustal deformation, it is not clear that it is valuable for detecting deformation precursors. If these are caused by slip over relatively small areas, a properly-sited obser-vatory instrument would be much more sensitive to the resulting deformation than a long-base geodetic measurement. In the absence of theoretical support for particular length scales of precursors I shall simply assume that both types of measurement are equally sensitive.

Given these assumptions we may compare noise from the two groups of methods quantitatively. We do this by finding the range of periods over which one technique is less noisy than the other and hence, given our assumptions, better. In this analysis we con-sider the effect of varying the sampling of the geodetic data from very long periods to near-continuous recording. Because the errors of two classes of techniques are described differently, this analysis requires some calculation.

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3. Comparison Method
By our definition of an error model, the only noise in an absolute measurement is from uncorrelated errors, with standard deviation rr a. If the true ground deformation at the time of the i-th observation is <?,, the observations are

I &l CL

The result of a series of relative measurements, on the other hand, is

<?. Q. + ft. t Jt t

where ni is the correlated noise contribution to the i-th observation. While rr a is just a number, ni is a time series and requires a more sophisticated description. Because it is a stationary sequence, it may be described by its one-sided power spectral density P(f). The spectra from high-quality tiltmeters and strainmeters have the form

P(f) = KIT (1)

for / < 10~ 4 Hz, with a roughly equal to 2. For observed data this is of course the spectrum of the observations <?,-, which include both measurement noise n- and ground deformation </,, but it must set an upper limit on the spectrum of n.

In order to compare the two measurement techniques we must express the errors (or noise) in the same way; adopting the more general description, we ask what P(f) is for the absolute methods. If we suppose the measurement to be repeated at regular intervals Ai, the independence of errors implies that the spectrum is flat up to the Nyquist fre-quency /n. Calling this level F0, we have by Parseval's theorem that

fn

/ P0 df = «l (2)
o

and hence P0 =&%tn where tn is the Nyquist period. The situation is then as shown in Figure 1: the noise spectrum for absolute measurements is flat and crosses the noise spec-trum for relative measurements at a period tm where

'» = RrH (3)

V Jtv)

For periods longer than tm the absolute methods are less noisy than the relative ones; for periods from tm to tn the reverse is true; and for periods less than tn the absolute methods are undersampled. There is not necessarily anything wrong with making tn » tm; this may be quite appropriate in terms of adequate sampling of large signals. However, when this is so, there will be a large period range over which relative measurements would have a lower signal-to-noise ratio. It is important to note that averaging or other lowpass filtering of the absolute measurements does not affect £m, but merely keeps it the same while increasing the sample interval.

4. Comparison Results
For sample computations of tm we need only approximate correct error models. For absolute measurements we take a priori estimates. The error (in strain) for baseline measurements (EDM or space techniques) has been taken to be

a ~~~ /) I /* / a 0 ~\- CL,

where L is the distance. For tilt from leveling the error is taken to be

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< = d/L

For single-wavelength EDM we use the error found by Savage and Prescott (1973): a scale uncertainty of 2-10~ 7 and a 3 mm centering error. For multiple-wavelength EDM we use the results of Linker et al, (1986): a scale uncertainty of 1.2-10~ 7 and a 0.7 mm centering error. For the "space techniques" we assume no scale error and a 1 cm standard deviation. In terms of the expressions above these become:

Method b
4-10" 14
1.5-HT 14

0

c(m2)
10~ 5

5-HT8

1(T4

EDM (1 color)
EDM (2 color)
Space

For first-order leveling we use a standard error of 0.5 mm/kmI/2 or d = 2.5-10" 10. For the spectra from relative measurements we use the data from the laser strain-meters and long fluid tiltmeter at Pinon Flat Observatory (Agnew, 1986). With / in Hz the coefficients are:

Quantity

K
a.

Tilt Strain
10~ 23 8-10' 27

2 2.5

Given these numbers we can compute tables of tm as a function of the geodetic baseline L and survey repeat time A£; these are shown in the tables below. In all tables L is in km, A£ and tm in days. Again, at periods longer than tm geodetic surveys outperform relative measurements; at shorter periods, the reverse. Leveling, unless done simultaneously by several crews, has limits to the distance that can be covered in a given time. For dis-tances and repeat times that could not be achieved by a single crew, the value of tm is printed in italics.

Period (Days) At Which Geodetic Surveys Outperform Strainmeters and Tiltmeters

Single Wavelength EDM vs. Strainmeter Two Wavelength EDM vs. Strainmeter £ (days)
15 50

(days)
150 500 .5 1.5 15 50

L (km)
3
10
30
50

1263
544
363
342

1959
844
564
531

3172
1366
912
860

4922
2119
1416
1334

7967
3430
2292
2159

L (km)
1
3
5
10

159
100
93
90

247
156
145
139

400
252
234
225

621
392
363
350

1005
634
588
566

"Space Techniques" vs. Strainmeter
(days)
15 50 150 500

L (km)
10
30
100
300
1000

1193
496
189
79
30

1852
769
293
122
47

2998
1245
475
197
75

4652
1932
737
306
117

7531
3127
1193
496
189

300
1000

Leveling vs. Tiltmeter
(days)
15 50 150

98 170 311 538

500

L (km)
10
30
100

538
310
170

932
538
295

1701
982
538

2946
1701
982

5379
2946
1701

54 98 170 295

982
538

841

We see from these tables that there is a considerable period range over which good relative measurements are quieter than absolute ones, even assuming frequent surveys. This is less true if the instruments are bad; for example, wire strainmeters are about 100 times noisier than laser strainmeters, so that to compare geodetic methods with such an instrument we should divide all values of tm by 10. For some of the longer times given, we are of course extrapolating the spectrum of relative measurements beyond the period range observed, but the conclusion will remain valid unless that spectrum rises more rapidly than the extrapolation. I know of no reason why this should be so.

5. Complications

The discussion above shows that we may increase the usefulness of absolute measure-ments for precursor detection by making them more often. There are two reasons for this. First, as A< is reduced, so is <n, so that higher-frequency signals would be measured without aliasing. Also, by equation (2), P 0 will fall as tn increases, causing tm to decrease (equation 3). For our ideal model we could make P 0 arbitrarily small and tm arbitrarily large by decreasing A< enough, but in practice this is unlikely to work, since it depends on the assumption that the errors <r a remain uncorrelated no matter how closely the observa-tions are spaced. More reasonably, we should assume that the error in absolute measure-ments are correlated only at time intervals less than some value te; this means (roughly) that the spectrum is at best flat up to a frequency f c =te l, and falls at higher frequen-cies (Figure 1). If we make Ad so small that tn < tc, the level of the flat part of the spec-trum will not depend on tni and neither will the crossover time tm] in this case only improvements in the measurement technique can reduce tm. Setting te = 0, as was done in the calculations above, biases the results in favor of the absolute measurements.

Since it is only recently that absolute geodetic measurements have been made fre-quently enough to look for fluctuations, the possibility of correlated errors has been little examined. A recent study by Savage et al. (1986) examines evidence of correlation in single-wavelength EDM measurements. They suggest, on a priori grounds, that the total error of measurement has two components: one caused by "systematic" errors in meteorological measurements that vary from survey to survey, and one that applies to pairs of measurements taken within each survey. There would thus be a correlation time corresponding to the mean time between changes of the meteorological sensors, presum-ably a few weeks.

Also, at the longest periods any deformation measurement would be expected to show the same divergent (and hence correlated) behavior as "relative" measurements because of monument instability. Wyatt (1982) showed that even high-quality near-surface monuments had horizontal displacements whose power spectrum had the form (1), with K = 6-10 23 and a = 2.75 (for a spectrum in m2 /Hz). By our earlier argument, this instability will become significant for times longer than
i/«

K

for a measurement over a baselength L. Application of the error models for <r a shows that this time is usually very long; e.g., for a 10-km line measured twice a year with single-wavelength EDM, monument instability contributes as much noise as measurement error only for periods greater than 30 years. This is in consonance with the lack of corre-lation found by Savage et al. (1986) in such data. On the other hand, for daily 2- wavelength EDM measurements over a 3-km line, monument instability would begin to dominate for periods longer than 480 days, and so might be important in evaluating

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precursor detection. While there might also be long-term divergences from any instability in the instruments, but these are probably not significant in comparison with monument motion.

6. Conclusions

A simple analysis shows that so far as noise levels alone are concerned, there is a wide frequency range over which the best relative measurements will do better than the best absolute ones. At present, even frequently repeated absolute measurements will be out-performed by good relative measurements at periods of a few months and less, though the scope for improvement of absolute measurements will probably change this in the future. By giving a general way to describe the errors from repeated absolute measure-ments, the discussion here allows us to decide if specified precursors would best be seen by particular methods.
The preceding discussion has assumed rather idealized models for the errors of some techniques. It is important to note that the procedure described here should be useful whatever the error model; as the dependence on distance and frequency become more com-plex no simple formula will give tm, but it can easily be found numerically, or indeed graphically.

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Deformation Measurement Spectra

log P

Geodetic Survey noise
Sampled
Continuous

Observatory -
based
Instrument noise

log f

m log t

tc

Figure 1

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RETARDATIONS IN FAULT CREEP RATES BEFORE LOCAL MODERATE EARTHQUAKES
ALONG THE SAN ANDREAS FAULT SYSTEM,
CENTRAL CALIFORNIA

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ABSTRACT

Records of shallow aseismic slip (fault creep) obtained along parts of the San Andreas and Calaveras faults in central California demonstrate that significant changes in creep rates often have been associated with local moderate earthquakes. An immediate post-earthquake increase followed by gradual, long-term decay back to a previous background rate is generally the most obvious earthquake effect on fault creep. This phenomenon, identified as aseismic afterslip, usually is characterized by above-average creep rates for several months to a few years. In several cases, minor step-like movements, called coseismic slip events, have occurred at or near the times of mainshocks. One extreme case of coseismic slip, recorded at Cienega Winery on the San Andreas fault 17.5 km southeast of San Juan Bautista, consisted of 11 mm of sudden displacement coincident with earthquakes of M 5.3 and M 5.2 that occurred 2.5 minutes apart on 9 April 1961. At least one of these shocks originated on the main fault beneath the winery. Creep activity subsequently stopped at the winery for 19 months, then gradually returned to a nearly steady rate slightly below the previous long-term average.
The phenomena mentioned above can be explained in terms of simple models consisting of relatively weak material along shallow reaches of the fault responding to changes in load imposed by sudden slip within the underlying seismogenic zone. In addition to coseismic slip and afterslip phenomena, however, pre-earthquake retardations in creep rates also have been observed. Onsets of significant, persistent decreases in creep rates have occurred at several sites about 18 months before the times of moderate earthquakes. A 44- month retardation before the 1979 M-5.9 Coyote Reservoir earthquake on the Calaveras fault was recorded at the Shore Road creepmeter site about 10 km northwest of Hollister. Creep retardation on the San Andreas fault near San Juan Bautista has been evident in records from one creepmeter site for the past 4 years. Retardations with durations of 25 and 19 months also occurred at Shore Road before the 1974 and 1984 earthquakes of M 5.2 and M 6.2, respectively.
Although creep retardation remains poorly understood, several possible explanations have been discussed previously. 1) Certain onsets of apparent creep retardation may be explained as abrupt terminations of afterslip from previous moderate mainshocks. 2) Retardations may be related to significant decreases in the rate of aseismic slip occurring beneath or within the underlying seismogenic zone. Such decreases may be caused by

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changes in local conditions related to growth of asperities, strain hardening, or dilatancy, etc., or perhaps by passage of stress waves or other fluctuations in driving stresses. 3) Finally, creep rates may be lowered (or increased) by stresses imposed on the fault by seismic or aseismic slip on neighboring faults. In addition to causing creep-rate increases or retardations, such fault interactions occasionally may trigger earthquakes.
Regardless of the actual mechanisms involved and the current lack of understanding of creep retardation, it appears that shallow fault creep is sensitive to local and regional effects that promote or accompany final preparation stages leading to moderate earthquakes. A strategy for more complete monitoring of fault creep therefore should be assigned a higher priority in our continuing efforts to test various hypotheses concerning the mechanical relations between seismic and aseismic slip.

INTRODUCTION

Evidence for creep activity on the San Andreas fault was discovered in 1956 during an investigation of damage within several structures at the Cienega Winery 17.5 km southeast of San Juan Bautista, California (Figure la; Steinbrugge and others, 1960). Occasional measurements of offsets between pairs of reference points established within the main winery building at the time of the initial investigation have continued over the years. Efforts to obtain continuous creep records in the building were begun in 1958 (Tocher and others, 1960), and at least one continuously recording creepmeter has been maintained at the winery ever since. The long-term average slip rate from combined offset readings and instrumental records at Cienega Winery is 12.3 mm/yr for the period 1948 to present (Figures 2, 3a).
The discovery that fault creep constitutes the dominant mode of shallow slip along the San Andreas fault between San Juan Bautista and Cholame was made during 1966 and 1967 in the course of detailed photo-geologic and field mapping (Brown, 1970). Preliminary compilations of the fault-trace map and notes concerning field observations of crack development and offset features facilitated selection of sites for construction of 25 alinement arrays distributed along the creep-active trace of the San Andreas fault from a point 16 km northwest of San Juan Bautista to a point 10 km southeast of Cholame (Figure la). This effort was begun in 1967 and was temporarily halted in 1975 (Burford and Harsh, 1980). Many of the original 25 alinement arrays are intact and several are still being surveyed. Several new arrays have been added since 1976, especially near Parkfield. Following the installation of the original rod and wire creepmeters in 1968 and 1969 (Schulz and others, 1982), additional sites for installation of creepmeters were selected both on the basis of alinement-array results and data from other fault-offset features (Figures Ib, c). Alinement-array and creepmeter results have contributed greatly to our knowledge of long-term distribution of shallow slip rates along the creep-active segment of the San Andreas fault (Figure 2). From these and other data, it has been suggested that the creep-active segment is divisible into three distinct sections (Figures la, 2); 1) a 55-km-long central section characterized by shallow creep that approximately matches the lateral movement between the adjacent major crustal blocks, 2) a northwest transitional section characterized by a northwestward decrease in average creep rate at about 0.3 mm/yr/km, and 3) a southeast transitional section (Parkfield-Cholame)

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characterized by southeastward decrease in average creep rate at about 0.5 mm/yr/km, if only during the latest interseismic period (1967-present; Figure 2). The central section contains a short segment near Monarch Peak with persistently lower creep rates than the average for adjacent sites, a feature associated with enhanced seismicity apparently related to local complexity in the fault zone (Burford and Harsh, 1980; Rymer and others, 1984). Creep records from Cienega Winery and other creepmeters installed since 1968 along the northwest transitional section and on the adjacent Calaveras fault (Figure Ib) show that creep rates are strongly influenced by local moderate earthquakes (3.9 < ML < 5.4). Additional evidence for correspondence between creep-rate fluctuations and moderate earthquakes is provided by records from creepmeters near Parkfield (Figure Ic). An increase in local creep rates for several months following moderate earthquakes is the most noticeable effect. Generally, post-earthquake accelerated creep shows a gradual decay in rate back to approximately the pre-earthquake long-term average. Of greater interest in the search for possible earthquake precursors are possible pre-earthquake retardations in creep rates. The purpose of this paper is to examine several examples of creep retardation and to discuss possible mechanisms concerning relations between seismic and aseismic slip.

SPATIAL RELATION BETWEEN SEISMIC AND ASEISMIC SLIP

There is a remarkable correspondence between the creep-rate profile and the distribu-tion of minor earthquakes along the creep-active segment of the San Andreas fault (Fig-ure 2). The 8-year detailed record of microseismicity represented by the lower curve (1977 through 1984) shows that minor earthquakes are concentrated along the northwest transi-tional section and at the northwest end of the southeast transitional section. Earthquakes are relatively scarce along the central, high-creep-rate section except for slightly higher ac-tivity in the vicinity of Monarch Peak, and the record of historic seismicity demonstrates that earthquakes of magnitude 4.0 or greater are rare within or near the central section, whereas earthquakes of this class occur frequently along the northwest and southeast tran-sitional sections in association with large numbers of microearthquakes (Ellsworth, 1975; Townley and Alien, 1939; Wesson and others, 1973). Average slip rates derived from offsets of older cultural features that have survived 2 or more characteristic Parkfield earthquakes are significantly higher than rates established by post-1967 alinement-array measurements (Figure 2). Thus it appears that the larger moderate earthquakes that occur about once every 22 years at Parkfield have a much more distinct effect on the accumulation of slip along the main fault than do more frequent but smaller moderate earthquakes along the northwest transitional section.

SAN JUAN BAUTISTA AREA (NORTHERN NETWORK) RECORDS

The distribution of creepmeter sites on the San Andreas and Calaveras faults near San Juan Bautista is shown in Figure Ib. With the exception of the record for Cienega Winery, which begins in 1958, creep records for this study begin at various times between 1968 and 1973. The detrended record for Cienega Winery through 1976 and detrended records for the rest of the northern network from 1968 through 1975 are shown in Figures 3a and 4a, respectively. More complete detrended records for the northern network from 1972/73 through mid-1986 are shown in Figure 4b. The principal examples of creep

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retardation in these records will be discussed in order of occurrence; 1) retardations and other effects associated with moderate earthquakes near Cienega Winery in 1959, 1960, and 1961, 2) retardations associated with the 1971-73 sequence of moderate earthquakes (Bear Valley sequence) along the northern transitional section, 3) retardations at Shore Road on the Calaveras fault associated with moderate shocks in 1974, 1979, and 1984, 4) a 46- month retardation near San Juan Bautista from 1976 through 1979 associated with a local sequence of moderate earthquakes during 1980 and 1981, and 5) recent retardations with onsets during mid to late 1982 at San Juan Bautista (XSJ2), Harris Ranch (XHR1,2), and Lewis Ranch (XFLl). The approximate locations of these and other relevant earthquakes are shown in Figure 5.
Composite record, Cienega Winery. The detrended composite record of shallow aseismic slip obtained at Cienega Winery from 1948 to 1976 demonstrates that slip at this site is normally stable at 12.3 mm/yr (Figure 3a). Times of major exceptions to this general condition are closely associated with moderate earthquakes concentrated in two distinct periods, the first between early 1959 and 1961, and the second between late 1971 and early 1973.
There is no detailed record of creep accumulation at the winery from time of construction until the time of initial investigations in April 1956 (Steinbrugge and others, 1960). All that can be said about the first 8 years of the record is that slip accumulated at an average rate slightly less than 13 mm/yr. From the record of repeated offset measurements obtained from 1956 through 1958, it appears that the average rate held approximately steady until about mid-October 1957. Creep subsequently accumulated at significantly lower rates varying between 7 and 11 mm/yr until early 1959. The surge of creep activity recorded during 1959 (about 20 mm) apparently was related to increasing seismic activity in the surrounding region (Figure 5a). The record of creep activity at the winery for 1959 is unique for three reasons: 1) About 11.3 mm of slip accumulated between the first of the year and April 10. A double creep event that began on April 2 accounts for 3.6 mm of that total. 2) A distinct 7-month-long retardation followed the early 1959 surge of creep activity. 3) The 7-month retardation ended with a single creep event of 6.0 mm amplitude that started on November 22. Except for the 1961 coseismic step, this is the single largest event on record for the winery site to date.
In summary, it appears from the early winery record that a 15-month creep retardation that began in 1957 ended early in 1959 with onset of a 4-month period of accelerated creep, associated with a possible increase in regional seismicity. Preceded by a possible short-term creep retardation (Figure 3a), the 6-mm creep event of November 1959 occurred 2 months before the the M-5.1 earthquake of 20 January 1960.
The source region for the 1960 mainshock was on the main fault beneath the winery (Figures 3b, 5a). A coseismic offset of 3 mm accompanied this earthquake (Tocher and others, 1960). From that time through the remainder of 1960, creep continued at a high rate (initially 20 mm/yr) with subsequent gradual rate decay to 13 mm/yr by the beginning of 1961, a pattern subsequently recognized as characteristic of shallow, aseismic afterslip (Nason, 1969; Burford, 1972). Following the decay in afterslip, creep activity was renewed in March 1961 with the onset of a 4.5-mm event (Figure 3a). Thus it appears that the pair of strong earthquakes that occured 2.5 minutes apart near the winery on 10 April 1961

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were preceded by 1) the January 1960 earthquake of M 5.1 on the adjacent fault segment to the northwest, 2) gradually decaying afterslip from the January 1960 earthquake that continued through about January, 1961, and 3) a creep event of 4.5-mm amplitude that occurred during 23-24 March 1961, 18 days before the double-mainshock event.
A coseismic step of 11-mm amplitude was recorded at the winery during the double-mainshock sequence (M 5.3 and 5.2; Figure 3a, b). From field evidence and the location determined for the first of these two mainshocks, it appears that at least one, and probably both of the shocks originated on the main San Andreas fault beneath the winery on the southeast side of the source region of the 1960 M-5.1 earthquake. Arrival times for P and S waves from the second earthquake were masked by signals from the first, resulting in considerable uncertainty in the location of the second event.
Creep activity was essentially arrested at the winery site for 19 months following the double mainshocks of April, 1961 (Figure 3a). Only about 1 mm of creep accumulated, and there were no distinct events between mid-April 1961 and the beginning of a 1.2-mm creep event recorded on 11 November 1962. The creep rate subsequently remained low at 7 mm/yr until mid-1964. During the last half of 1964, the creep rate gradually increased, reaching a stable value of 10.5 mm/yr by the beginning of 1965 that persisted through about mid-1970.
Except for postulated source areas for the M-5.1, M-5.3, and M-5.2 mainshocks of 1960-61, the source areas for moderate mainshocks indicated in Figure 3b are based on distributions of hypocenters for related aftershocks recorded for 72-hour intervals after each of the mainshocks (R. Wesson, oral comm., 1974). It appears that a strong case can be made for placing the 1960 and 1961 mainshock source areas in the seismic gap defined by the 1971-73 earthquake sequence and the pattern of subsequent microseismicity (Figures 2 and 3b). Epicenters determined for the 1960 and the first of the 1961 mainshocks suggest unilateral rupture propagations from northwest to southeast through the postulated source
areas.

Northern network creep records from 1968 to present. From 5 August 1970 through June 1971, creep accumulated at Cienega Winery at only 6 mm/yr, approximately 51% below the previous long-term average (Figure 3a). This retardation lasted about 12 months until the onset of a 5.5-mm creep event on 27 June 1971. The year 1971 closed with a 4.6- mm event that accompanied the first pair of magnitude 4 events of the 1971-73 moderate earthquake sequence (29 December 1971, ML 4.0 and ML 4.1, about 1.25 hours apart). The December 29 "Limekiln Road North" events occurred on the San Andreas fault about 8 km southeast of the winery (Figures 3b, 5b).
The mid-1970 retardation onset at the winery was relatively weak but nevertheless shows clearly in the record because of the previous 6 years of remarkably stable creep rate (Figures 3a and 4a). Factors that help explain a low-amplitude retardation are 1) the earthquakes that occurred during the Bear Valley sequence of 1971-73 were relatively small near the winery (ML 4.0 to 4.2; Figures 3b, 5b), and 2) the winery occupies the the approximate center of the southeast half of a 15-km-long quiet gap in the seismicity pattern documented since the late 1960's. This gap is apparently related to the aforementioned larger moderate mainshocks in 1960 and 1961 (Figure 3b; Burford and Harsh, 1980), and

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it appears that creep activity within the gap has been relatively immune to the influences of moderate earthquakes along adjacent sections of the fault since 1961.
A third possible retardation appears in the winery creep record during late 1973 through 1974 (Figure 3a). Although similar in amplitude and duration to the 1970-71 example, there were no local moderate earthquakes late in 1974 or early in 1975 that can be associated with the 1974 retardation. Coming as it did after the 1971-73 sequence of moderate earthquakes, it is perhaps a reflection of the previous period of instability rather than a response to continuing instability of the type that might have led to additional earthquakes. Taken at face value, the 1973-74 retardation recorded at the winery is a false alarm.

Unfortunately, creepmeters at several sites near Cienega Winery were not installed until 1968 at the earliest and most were not functional until mid-1969 or later (Figure 4a, b). Consequently, records at all sites other than the winery begin very late before mid-1970 and therefore provide inadequate baseline information for clear recognition of possible retardation onsets preceding the 1971-73 Bear Valley earthquake sequence. Nevertheless, it appears that by correspondence with the more complete records from the winery, a good case for late 1969 to mid-1970 onsets of retardation at HRS3, XMRl, and MRB1 creepmeter sites can be made. The retardation was particularly strong at the Melendy Ranch site XMRl. Creep rates were also relatively low at SJN1 and XSJ1 near San Juan Bautista, but clear retardation onsets are not contained in the records, perhaps owing to late starts for the monitoring. Low rates at each of the sites ended approximately at the times of the nearest of the moderate earthquakes in the 1971-73 Bear Valley sequence (Figure 4b).
Creep retardations on the Calaveras fault at Shore Road. The original creepmeter
at Shore Road (SHR1) on the Calaveras fault north of Hollister was established in 1971 (Figure 4c). Between May 1971 and late December 1972, the local creep rate on the Calaveras fault was 11 mm/yr, but during late December 1972, the creep rate decreased to 3 mm/yr and remained low until the onset of a 9-min creep event on 14 September 1974 (Figures 4b, c). The M-5.2 earthquake of 28 November 1974 occurred on the nearby Busch fault soon after the end of the 1972-74 retardation. The epicenter was about 6 km southwest of the creepmeter near the intersection of the Sargent and Busch faults (Figure 5c). As a direct result of the large creep event in September 1974 and a second, 8-mm event on 8 July 1975, the creep rate at SHR1 was relatively high at 18 mrn/yr during the 1-year period containing the 1974 mainshock.
A period of creep retardation with one of the greatest durations on record to date started at Shore Road sometime between late 1975 and early 1976 (Figure 4b). By mid-1976, creep was essentially arrested at SHR1 and remained so until late 1977 (Figure 4c). Between late 1977 and the time of the Coyote Reservoir earthquake of 6 August 1979 (M 5.9), creepmeter SHR1 recorded a very slight, gradual rise in creep rate (<1 mm/yr, at most). The creep rate remained very low until a 4.2-mm coseismic step accompanied the M-5.9 mainshock. The mainshock epicenter was 15 km northwest of Shore Road (Figure 5c), but the aftershock zone extended southeastward to the creepmeter site. The coseismic response at SHRl was immediately followed by a cluster of creep events over

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the next 5 days (Raleigh and others, 1979). The total for combined coseismic slip and immediate afterslip was about 13 mm.
Following the 1979 mainshock and corresponding afterslip at SHRl, the creep rate stabilized at 13 mm/yr, well above the previous average, until late 1982 (Figure 4b, c). The 1982 decrease in rate is associated with the occurrence of a moderate earthquake (M 4.6, 11 August) on the San Andreas fault near creepmeter XFL1 (Figures 4b, 5c), about 35 km southeast of Shore Road. Subsequently, a relatively low, 2 mm/yr creep rate persisted at Shore Road until the time of the M-6.2 Morgan Hill earthquake on 24 April 1984. The epicenter for the Morgan Hill earthquake was 50 km northwest of Shore Road on the Calaveras fault but the aftershock zone extended southeastward to within about 21 km of the creepmeter (Schulz, 1985b).
A small coseismic step (about 0.1 mm) at Shore Road accompanied the 1984 mainshock. Within the next 18 hours, afterslip of 12.9 mm occurred (Schulz, 1985a), and the rate remained high at about 18 mm/yr, at least througa the end of the SHRl record which was terminated by the complete failure of the instrument during the 26 January 1986 Tres Pinos earthquake 38 km to the southeast (Figure 5c). The replacement instrument, XSH1, became operational during May, 1986. To date, little activity has been recorded by the new instrument.
Creep retardation near San Juan Bautista, 1976-1980. A comparison of detrended records from SHRl on the Calaveras fault and from XSJ2 on the San Andreas fault about 12 km to the southwest (Figure 4b) shows long-period fluctuations in creep rate that are nearly coherent at these sites. Conformity between these records is improved when an apparent 9-month delay at XSJ2 with respect to SHRl is considered. However, the period of low creep rate at XSJ2 between early 1976 and early 1981 shows little if any response to the 1979 Coyote Reservoir earthquake, in contrast to the record from SHRl. Instead, the first sign of gradual recovery toward normal creep activity came after a local moderate eathquake occurred nearly beneath the instrument (M 4.7, 13 April 1980, about 2 km SE of XSJ2; Figures 4d and 5c). A second moderate earthquake of ML 4.5, centered about 10 km northwest of XSJ2, occurred on 7 January 1981, and within 4 months, the creep rate increased dramatically and remained high at 18 mm/yr until the August, 1982 onset of the most recent retardation. Responses from other nearby creepmeter sites were minimal (Figures 4b, d).
Recent retardations with mid- to late-1982 onsets at XSJ2, XHRl/2, and XFL1. Of considerable interest at present are the onsets of apparent creep retardations at XSJ2, XHRl, and XFL1 at about the time of the M-4.6 Stone Canyon earthquake of 11 August 1982. Although onsets of these retardations closely coincided with the onset of the latest retardation at Shore Road, the low rates on the San Andreas fault did not end at the time of the Morgan Hill earthquake of May 1984. Despite several minor creep events at XSJ2, XHR1/2, and XFLl and the lack of significant retardation at Cienega Winery, creep rates remain low at present, especially at XSJ2. The invar wire at creepmeter XHRl broke during August 1984 and was not replaced until early April 1985. The zero value was lost, hence the new site name XHR2. The new rate recorded by XHR2 since April 1985 is 7.2 mm/yr, only slightly below the previous long-term average of 7.5 mm/yr. Thus,

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retardation at XHR2 may be over, even though the average rate since early 1983 is still 2 mm/yr below the previous long-term average.
The creep rate at XFLl remained low (28% below long-term average and 49% below the trend established during 1980 through August 1982) until July, 1986, one month after the nearby earthquake of ML = 4.8 on May 31. The limited data available since the apparent increase to 8.3 mm/yr preclude a definite conclusion that the recent retardation ended with the 1986 earthquake, but this is a strong possibility (Figure 4d). The May 1986 earthquake apparently resulted from a repeat breakage of the same source area as for the M-4.9 mainshock of 4 September 1972 (Figures 3b and 5b, c; W.L. Ellsworth, oral comm., 1986).

PARKFIELD-CHOLAME AREA (SOUTHERN NETWORK) RECORDS

After decay of afterslip from the 1966 Parkfield earthquake sequence, creep at Carr Ranch (CRRl), 11 km southeast of Parkfield, stabilized at 10 mm/yr and remained nearly constant until 1976 (Figure 6). In contrast to the regularity at CRRl, other instruments installed in the Parkfield area between 1969 and 1985 show greater variability in creep rate and in character of slip accumulation than displayed by the creepmeter sites in the northern network. Presumably this is due to greater complexity in local fault geometry within the fault zone in the vicinity of Parkfield. Earthquakes within the surrounding region, including shocks on major thrust faults 25 to 35 km to the northeast, also have exerted a strong influence on creep activity at some of the creepmeter sites in the vicinity of Parkfield. For instance, a pair of moderate earthquakes in 1975 (M 4.9, 3 August, near Coalinga; M 4.8, 13 September, about 18 km NW of Parkfield) mark the approximate time of rate increases at XSC1 and at XGH1 and a possible decrease at XDRl. The decrease at CRRl shows a very gradual onset sometime after early 1976 without any obvious response to the mainshocks in 1975. Although perhaps related to the 1975 earthquake, obvious retardation at CRRl did not occur until late 1977, accompanied by a slight rate decrease at XSC1.

Recent retardations at XMM1, XPK1, WRKl, and XGHl are clearly associated with the effects of the M-6.7 earthquake that occurred on 2 May 1983 near Coalinga. Obvious retardation ended at XPK1 at the time of the Kettleman Hills earthquake (5 August 1985). These effects are discussed in a companion paper (Simpson and others, this volume) and will not be considered in detail here. It remains to be seen what relation the recent

and continuing creep retardations near Parkfield will have to the anticipated, impending recurrence of a characteristic Parkfield earthquake.

DISCUSSION

Coseismic steps and afterslip responses are readily understandable as effects of sudden changes in load on relatively weak material along the shallow reaches of the fault surface. The extreme weakness of shallow fault gouge may cause strains to be concentrated across the fault trace at certain localities, thereby promoting high sensitivity to regional stress changes at favorably located sites (King and others, 1977; Bilham and Beavan, 1979; Burford and Schulz, 1984; Mavko and others, 1985).

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Pre-earthquake retardations recorded at various creepmeter sites are not as well understood as coseismic and afterslip effects. In most of the documented cases it is not clear whether surface creep retardation is a direct response to a decrease in slip rate within the underlying seismogenic zone or merely a surficial response to regional stress changes from mainshocks on other parts of the fault system. The latter hypothesis is attractive because of several examples of retardation onset near the times of such shocks. The modeling work of Mavko (1982) for fault interactions near Hollister also supports this explanation. However, observations of concurrent seismic quiescence accompanying recent creep retardations near San Juan Bautista and Bear Valley suggest that changes in behavior within the underlying source patches of the seismogenic zone may be responsible for surface creep retardations (Wyss and Burford, 1985a; Wyss and Burford, 1985b). This explanation does not rule out the likelihood that both seismic quiescence and creep retardation can be triggered. Moreover, it leaves open the possibility that in certain cases, spontaneous local changes in slip resistance along the fault plane may control the onsets of both of these phenomena. The 1977 retardation onset at CRRl creepmeter near Parkfield may be an example of abrupt termination of slip at the end of a prolonged afterslip response (Figure 6). Other possible examples are contained in 1973 to 1980 records from creepmeters HRS3 and XFL1 (Figure 4d). Conversly, the record from CRRl may indicate that slip has merely transfered to another trace. However, the nearly concurrent decrease in creep rate at XSCl from well above to slightly below the approximate long-term average and the nearly simultaneous rate increase at WRK1 (Figure 6) suggest other than strictly local causes for the strong decrease at CRRl.

Retardations at Shore Road may have been promoted by fault interactions as suggested by Mavko (1982), but relations between duration of retardation and both magnitude and relative remoteness of source areas for the 1979 Coyote Reservoir and 1984 Morgan Hill earthquakes suggest onsets of slip-rate decrease within those source areas before the conditions necessary for mainshock occurrence were reached. Both of these mainshocks also may have been preceded by seismic quiescence (Max Wyss and Ted Habermann, oral communication, 1985). As mentioned above, there is circumstantial evidence that the 1982-84 retardation may have been triggered by the M-4.6 earthquake of 11 August 1982 (Figure 4b, d).
Preliminary results of a fresh investigation of the effects of fault interaction are presented and discussed in a companion paper (Simpson and others, this volume). Many of the observations discussed above support the idea that fault interactions are common and potentially very important causes of sudden changes in behavior of shallow reaches of the fault. It remains to be seen whether fault interactions focus on only the shallow slip regime or on the entire seismogenic zone plus adjacent reaches of the fault surface. In any case, the fact that the most obvious surface creep retardations have invariably ended at or slightly before the times of nearby mainshocks supports the hypothesis that the two phenomena are directly related. Current and future occurrences of persistent creep retardations thus may be used to identify target areas for study of relations between seismic and aseismic slip as well as other intermediate-term precursory phenomena.

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figure 1c
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Figure la. Map of San Andreas fault system in central California showing sites for 25 alinement arrays established between 1968 and 1975, as indicated by short dashed lines across the main fault trace. The creep-active segment of the main San Andreas fault between San Juan Bautista and Cholame can be divided into a high-slip-rate central section flanked by northwest and southeast transitional sections adjacent to the ends of currently locked segments that slipped during great earthquakes of 1857 and 1906 (Burford and Harsh, 1980). Geographic reference points provide ties to the slip-rate and seismicity profiles of Figure 2.

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121°37.10'

37°00.00'

121°04.30'

36°31.35'

HOLLISTER

SAN
JUAN
BAUTISTA .
HRS3, XHR1.2

CIENEGA

v* SCR1,3&
XMRT^\
MRB1
10 km

Figure Ib. Creepmeter locations along creep-active traces of the San Andreas fault system in the vicinity of San Juan Bautista (Northern Network). Site names in italics designate creepmeters that have been replaced or sites that have been abandoned.

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120°30.58'
35°56.35'

120°16.64 f

35°43.28'

^AXMM1

PARKFIELD

Figure Ic. Creepmeter locations along the southeast transitional section of the San Andreas fault in the vicinity of Parkfield (Southern Network).

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NW DISTANCE Ckm) SE Figure 2. Distribution of surface slip rates along the creep-active segment of the San Andreas fault in central California between San Juan Bautista and Cholame (slip-rate scale on left) in comparison with distribution of minor earthquakes recorded during the period 1977 to 1984 (ML > 1.5, cumulative-count scale on right). Slip-rate data from alinement surveys and from measurements of offset cultural features shown by small circles and squares, respectively (data from Burford and Harsh, 1980). Profile of number of minor earthquakes per 5-km section is given as continuous curve. Peak of minor earthquake production (about 6 shocks/km/year) is centered in southern Bear Valley, near Pinnacles National Monument, about 47 km southeast of San Juan Bautista. Subordinate peaks are centered near San Juan Bautista, Monarch Peak, and at a point between Bagby Ranch and Claassen Ranch sites about 13 km northwest of Parkfield.

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CIENEGA WINERY

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M4.2

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M4.5
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Figure 3a. Cumulative fault-creep record from Cienega Winery on the San Andreas
fault 17.5 km southeast of San Juan Bautista, California, 1948 through 1976. This composite record was assembled from measurements of progressive building offsets from April 1956 through mid-1958 and the results of continuous instrumental monitoring from mid-1958 through 1976. The 1948 zero point is based on building construction date (approximate time that concrete floor slabs were poured). Parts of the instrumental record from 1969 forward are included in Figures 4a and 4b. The raw record was detrended by extracting the long-term average slip rate of 12.3 mm/yr. Average creep rates before and after the effects of the 1960-61 earthquake sequence were 13 and 10.5 mm/yr, respectively (latter from mid-1964 through 1970).

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M5.1
1/20/60

M4.0,4.1
12/29/71

M4.8
10/3/72

M4.6
9/4/72

M5.2,5.3
4/9/61

M4.2
1/15/73 M5.0
2/24/72

0 10 20 30

FAULT LENGTH Ckm]

40 50

Figure 3b. Cross section along slip surface of San Andreas main fault from San Juan Bautista to Bear Valley showing relations between surface creepmeter sites and fault-plane source areas for moderate mainshocks that occurred during the 1960-61 and 1971- 73 earthquake sequences. Source areas for shocks in the 1971-73 sequence are based on the details of related aftershock hypocenters recorded for 3 days after each of the mainshocks (R. Wesson, oral comm., 1974). The postulated source areas for the 1960 and 1961 mainshocks are scaled approximately from comparable sources and magnitudes for the 1971-73 events. The 1960 and 1961 source areas are also suggested by relative hypocentral locations, field evidence for limited surface ruptures along the main fault trace, details of coseismic slip contained in the winery creepmeter records, and the persistent gap in minor earthquake distribution from the 1969 to present-day catalog. The gap in recent seismicity at Cienega Winery is thought to be a "shadow" of the 1960-61 sources.

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SHORE ROAD CUMULATIVE CREEP

Morgan Hill
earthquake

Coyote Lake
earthquake

1970 1971 1972 1973 1974 1975 1976 1977 1978 1979 1980 1981 1982 1983 1984

Figure 4c. Cumulative creep record from Shore Road instrument SHR1, mid-1971 through late 1984.

Figure 5. Epicenters of moderate earthquakes in the Bear Valley, Hollister, San Juan Bautista area: a) 1959-1961; b) 1971-1974; and c) 1974-1986.

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Summary of session on Laboratory and Crustal Deformation Measurements

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lull is reviewed recent laboratory observations on rock friction emphasi-zing time and rate dependent effects. These observations have led to a state variable description that has been remarkably successful in predicting whether slip under a variety of laboratory conditions will be stable or unstable. Because this constitutive description is suitable for surfaces that have slipped many times, it holds great promise for application to crustal faults, as suggested by lull is 1 review of the work of Tse and Rice [1986], and has the possibility of being a unifying concept in fault mechanics that can explain many observations.

Nevertheless, a number of difficulties must be overcome. Prominent among these are the scaling of the critical sliding distance that appears in the state variable formulation and a better understanding of the rate/state depen-dent friction relation when used between continuous elastic solids. lull is reviewed current work on scaling the distance parameter, which has focussed on understanding the relation of this length scale to surface roughness, but efforts have been hampered by a lack of detailed understanding of the micro-mechanical processes responsible for the observed phenomenology. Understanding of the rate/state dependent friction relation in elastic systems has been primarily based on studies of one degree-of-freedom systems which are crude models for faults in the earth's crust. Consequently, there is a need for better understanding of the response of continuous elastic solids contain-ing faults with rate/state dependent friction. Unfortunately, modelling of even simple two-dimensional fault geometries has been limited by computational
resources.

Three presentations described specific observations related to particular aspects of crustal deformation: block rotations (Nur), fault bends (King) and creep retardation preceding moderate earthquakes (Burford). In each case the complex geometry of the fault systems studied contrasted with the idealized ones for which realistic constitutive relations have been studied. Explanations of the observed behavior emphasized the dominant role of kinematics. Clearly, a great need exists, complementary to that for study of oversimplified fault models employing realistic constitutive relations, for understanding the kinematics of realistic fault geometries with simplified constitutive relations.

Two presentations discussed crustal deformation mesurements: Beavan reviewed a variety of techniques, emphasizing the need for more long-term data, better measurements of environment, and comparison of crustal deformation measurements with other precursors; Agnew focussed on the problem of noise in measurements and, more specifically, on the frequency of measure-ment needed to achieve comparable noise levels by different techniques. Both noted that there have been few if any convincing examples of precursors.

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However, because of inadequacies in both modelling and measurement, it is not possible to say whether there have been no precursors or they have not been detected. Although both presentations emphasized the unsatisfactory nature of results to date, they offered specific suggestions for improvement and it is promising that a clearer picture of what needs to be done is emerging. Whether the committment in time, manpower, and funding will be sufficient to accomplish this remains to be seen.

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PHYSICAL MODELS FOR EARTHQUAKE INSTABILITY AND PRECURSORY PROCESSES

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SUMMARY

In physical models of earthquake instability, the instability arises as a result of interaction of a fault constitutive relation with deformation of the surrounding material that occurs in response to remote tectonic loading. In contrast to kinematic models in which the fault slip is imposed it is calcu-lated in physical models and, consequently, these models are essential for understanding precursory processes. Furthermore, they can provide a framework for the interpretation of field observations and aid in the efficient deployment of observation networks.
Some kind of "weakening" behavior for the fault constitutive relation is required to produce an instability analogous to an earthquake. Two commonly employed idealizations are rate-independent slip weakening and rate-or-state dependent friction. In the former, shear stress decreases from a peak value T to a residual value T as slip increases. In the latter model, the shear stress associated with sliding at a constant slip rate (with constant environ-ment) evolves toward a steady-state value T S (V). Both models are capable of exhibiting instabilities in a sufficiently compliant system, the rate-or-state dependent models only if d-r /dV < 0, but the rate-or-state dependent models are more complex and exhibit a correspondingly richer variety of response. Moreover, they are capable of simulating rehealing and repeated events.
Both of these constitutive models are motivated by and consistent with laboratory observations. However, parameter values appropriate to field con-ditions are largely uncertain. In particular, both models involve a length scale for which field values are inferred to be several orders of magnitude larger than values measured in the laboratory. Whether this discrepancy is simply a matter of scaling or a reflection of fundamentally different processes in the field and laboratory is unknown.
When these constitutive models are employed on surfaces embedded in elas-
tic half-spaces or layers, possibly coupled to a viscoelastic substrate, the results are capable of simulating realistically some aspects of earthquake occurrence. However, the complexities are such that most computations have been done for two-dimensional models using rate-independent slip weakening. A considerable simplification results when most of the slipping portion of the fault is sliding at the residual friction level and the decrease of stress

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from T to T occurs in a region near the fault edge having a length much less than the overall crack length. In this case, the slip weakening model reduces to an elastic-brittle shear crack. That is, stresses at the fault edge are singular and advance of the slip region occurs when the energy released by virtual advance is equal to a critical value. This value can be identified as the area under the shear stress versus slip curve in excess of the residual value T . Three-dimensional calculations are rare but some progress has been made by use of the "line-spring" approximation. This procedure approximates the three-dimensional problem as the solution to two two-dimensional problems: one relates the variation of thickness averaged stress and slip along strike; the other calculates the depth variation at each section.
Common to all models is the prediction that earthquake instability is preceded by precursory slip. This precursory slip produces a departure of surface strain-rate from the background level. Near the epicenter of a moder-ate to large earthquake, the magnitude of this departure appears to be well within the range of current geodetic measurement accuracy and its duration is of the order of months to years. However, details depend on a variety of factors, including the detailed modelling of the constitutive relation near peak stress, coupling of elastic crust to the asthenosphere, and coupling of deformation with pore fluid diffusion.
Although the complexity and sophistication of these physical models have increased greatly in the last decade, much remains to be done. Accurate and efficient methods have not yet been employed for three dimensional calcu-lations of slip progression both along strike and up from depth. Partly as a result, there is yet little understanding of slip advance into or around strongly slip resistant zones. Nor has there been much modelling of bends and offsets, features that are thought to be important in rupture initiation and termination. More generally, effects due to perturbations in the geometry of the slip front have only begun to be investigated. Current calculations can simulate realistically the occurrence of a main event, but computational limitations prevent the modelling of smaller scale instabilities indicative of minor seismicity. Consequently, there has been only qualitative connection of predictions of physical models with observations of seismicity variations. Also, current modelling idealizes all inelastic deformation as occurring on a surface and, hence, provides limited insight into precursory observations that

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seem to require more widespread inelastic deformation. Effects of fault interaction also have not been investigated thoroughly.

INTRODUCTION

If data were more numerous and earthquakes more frequent, it is, perhaps, possible that earthquakes could be predicted purely on an empirical basis. This is, however, not the case: earthquakes that are large enough to possibly exhibit premonitory effects and that occur in heavily instrumented areas are infrequent. Consequently, purely empirical methods of earthquake prediction, no matter how promising, are viewed with skepticism until there is a good understanding of the physical basis for the method. Physical models can help to understand the conditions that give rise to earthquake instability and the processes that precede them. Moreover, they are the only means of exploring the consequences of particular aspects of the earthquake process in isolation from the complexities of competing mechanisms occurring in the earth. Also, physical models can aid in the strategic deployment of instruments and in the interpretation^of observations. Furthermore, physical models provide the only sensible basis for extrapolating laboratory observation to field conditions. Because there is no direct source of information about constitutive properties of materials in situ, information about material behavior mainly comes from laboratory observations at size and time scales that are vastly different from those in the field.
Physical models are those that incorporate plausible constitutive rela-
tions consistent with laboratory observations and compatible with the laws of mechanics. In these models, an instability, which can be interpretted as an earthquake, arises because of the geometry or the nature of the constitutive relation. In quasi-static models, in which inertia is not included explicitly, the instability generally takes the form that the slip rate becomes unbounded on a portion of the fault. In models which do include inertia the slip rate does not become unbounded but there is a transition from a situation in which inertia is negligible to one in which it is not. Physical models can be distinguished from kinematic models in which the distribution of subsurface slip is assumed or inferred. Kinematic models are particularly useful for interpreting data and have been extensively used to infer

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distributions of subsurface slip that are consistent with surface deformation measurements. However, these models are not able to predict whether further slip will occur and thus, are limited in their capability to illuminate the physical mechanisms and geometric conditions that lead to instability. Physical models have increased greatly in complexity and sophistication
in the last decade. These advances have been motivated primarily by improve-ments in constitutive descriptions based on laboratory observations and in geodetic measurements. An indication of the progress in understanding is the current general agreement that some kind of weakening behavior is needed for instability; at a previous USGS workshop on fault mechanics [Stuart et al., 1977] this was a point of contention. Despite the progress, there is still much to be accomplished. Models are simplistic in terms of the geometric complexities of actual fault systems. There are still gaps in the parameter ranges explored by laboratory observations and a general lack of understanding of how to extrapolate these observations to the field. Models have been limited, with a few exceptions, to two-dimensions and straight, continuous faults. Even with these simplifications, investigation of relevant ranges of parameters has been limited by computational resources.
This review of physical models is neither comprehensive nor thorough.
Instead, it concentrates on a few aspects of these models and, in parts, draws heavily on a recent more extensive review by Rice [1983], I begin by describing the constitutive relations that are typically employed in these models and the way in which they give rise to instabilities. Then, procedures for incorporating these relations into crustal scale fault models are described. After a brief outline of some of the results of these models, the paper concludes with a discussion of some of the limitations of these models and suggestions for further research.

CONSTITUTIVE MODELS

Most fault constitutive models relate the stress (traction) at a point on
the fault surface to the relative displacement at a point on the fault surface. An implicit assumption in such models is that essentially all inelastic deformation is confined to the neighborhood of the fault surface. Although earthquake faults are generally localized features, there may be

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instances in which inelastic deformation is more broadly distributed prior to instability. In such cases, an inelastic constitutive relation for bulk deformation [e.g., Rudnicki, 1977] is needed.
Because no ideal point measurements are possible, fault surface
consitutive models are inferred from laboratory experiments in which slip occurs essentially simultaneously over the entire fault surface. As mentioned already, there is a consensus that some kind of weakening behavior is required for instability. Two common idealizations, to be discussed in more detail below, are the rate-independent slip weakening model and the rate and state dependent friction.
The slip weakening model is illustrated in Figure 1. The shear stress T required for slip first increases to a peak T , then decreases to a residual value T after an amount of relative slip 6 (Figure Ib). Slip is generally accompanied by uplift or microcracking of the adjacent material [Barton, 1976; Teufel, 1981; Raleigh and Marone, 1986] (Figure Id). This dilatancy is typically small, corresponding to uplift equal to a few percent of the slip, but can be important if pore fluids are present [Rudnicki and Chen, 1986].

The peak stress T and the residual stress T depend on the effective p r r
compressive stress on the fault a-p (where a is the total compressive stress and p is the pore fluid pressure), the temperature, and, possibly, preparation of the surface, but are idealized as being independent of the rate of sliding (Figure Ic). The difference T - T is thought to first increase with depth in the earth but then to decrease, reflecting a transition to more ductile conditions.
Slip weakening curves have been observed in laboratory shear tests (e.g. Dieterich [1978], Barton [1972, 1973]) and have been inferred from observa-tions of postpeak deformation in axisymmetric compression tests on both initi-ally intact and sawcut specimens [Rice, 1980; Wong, 1982]. These models can be regarded as a generalization of the concept of static and kinetic friction: instead of an abrupt drop from the static to the kinetic value with the onset of slip, there is a transition reflected by the decrease of shear stress from peak to residual value. Although these models neglect rate-dependent effects to be discussed shortly and in the article by Tullis [1987, this volume], they are consistent with more elaborate rate-and-state dependent models under conditions of uniformly accelerated motion [Gu, 1984/85]. The primary

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shortcoming of these models is that they lack any mechanism for regaining strength after reduction to the residual level and, consequently, they are inadequate for simulating repeated events on the same fault segment. Nevertheless, the slip weakening models are a good approximation under many circumstances and, by comparison with the rate/state dependent models, have an appealing simplicity.
Experiments initiated by Dieterich [1978, 1979, 1981] (also, Ruina [1983, 1984] and Weeks and Tullis [1985]) have demonstrated a rate-dependence of friction that is not modelled in the slip weakening idealizations. The canonical experiment is shown schematically in Figure 2. Sliding at constant normal stress at a velocity V, gives rise to a steady-state value of resisting shear stress t ss (V 1). The sliding velocity is now suddenly increased to a velocity V«. The shear stress exhibits an instantaneous increase and then evolves to a new steady state value T ss (Vp) over a characteristic sliding distance, L. The new steady state shear stress level may be greater or less (as shown in Figure 2) than T (V,); that is, a? /8V may be positive or

negative. Laboratory results at room temperature generally yield 81 (V)/8V j o negative, but some results for adolescent surfaces and higher temperature yield positive values (e.g., see Tse and Rice [1986] for a summary). As will be discussed shortly, the sign of 81 (V)/8V is critical to the stability of small perturbations from steady-state sliding: instability is possible only

if 81 (V)/8V < 0 (velocity weakening). j O
Dieterich's [1978, 1979] original modelling of these experimental obser-vations has been reworked into a state variable description due to Ruina [1980, 1983]:

T = F(V, *.)

'l- G(V, ij,.) (i = 1, 2, ... N)

where the ty. are a set of variables that describe the state of the surface. For steady sliding, the i|». are constant and the steady state shear stress is

T SS = F(V, ^s) (2)

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The change in shear stress in response to the sudden change of velocity illustrated in Figure 2 is assumed to occur at constant values of the state variables associated with steady sliding at velocity V,. Thus, the evolution of the shear stress toward the new steady state level is associated with changes in the state variables. Equations (1) may depend additionally on the normal stress, temperature and environment.
One state variable is usually adequate to describe experimental results, although sometimes two are needed. If one is sufficient Rice [1983] (also, Rice and Tse [1986]) has shown that the formulation can be recast into a form that eliminates specific reference to the state variable:

dt

_ r dt L LT

where L is a length scale that is a measure of the slip distance needed to establish the new steady-state shear stress after a sudden change of velocity. Various minor modifications of this constitutive relation and appropriate values of the parameters that enter it have been widely discussed by a number of authors (also, see Tullis [1987]).
An important feature of both the slip weakening and the rate/state-
dependent relations is the appearance of a characteristic sliding distance. In the slip weakening relation, this is 6 , the slip needed to degrade the strength from peak to residual value. In the rate/state dependent model, this is the distance L needed to establish a new steady-state shear stress. In
C _A experiments, these distances are typically of the order of 10 to 10~ m. These characteristic distances tend to increase with surface roughness but the scaling to values appropriate to crustal faults is largely uncertain. Values _2
very much larger than observed in the laboratory, of the order of 10 to
10~ m, are compatible with inferred energy release rates for elastic-crack fault models with slip weakening [Rice, 1983], The extrapolation of labora-tory values for the critical distance to field situations would be aided by an understanding of the micromechanical processes that are responsible for the phenomenology. Presumably, the rate dependent version results, as Dieterich [1978, 1979] has argued, from competing processes of asperity deformation and remating of asperity contacts, but the details of this process are not understood. Current efforts at scaling have focussed on determining roughness

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values for faults in the field. However, there must be some question whether the micromechanical processes of sliding on well-defined surface or gouge zones in laboratory specimens are the same as those in the field where overall macroscopic slip may be accomplished by more complex processes of slip on en echelon surfaces linked by regions of tensile fractures.

INSTABILITY

Both the slip weakening and rate/state-dependent models give rise to instabilities that can be interpretted as earthquakes. These instabilities can be explained most easily in terms of a single degree-of-freedom spring-slider system (Figure 3a). The single degree of freedom system is most appropriate as a model for laboratory experiments in which slip occurs essentially simultaneously on the entire sliding surface, but can also be used as a crude model of an isolated fault (Figure 3b). In the latter case, the friction law is regarded as relating the average stress and slip on the fault. The effective stiffness of the surroundings, corresponding to the spring constant k, depends on the elastic constants and is inversely proportional to the fault length 2a. This idealization neglects effects due to the inhomogeneity of slip on the fault and the stress concentrations at the ends. Although the single degree-of-freedom system is a crude model of a fault, most of the results to date for the rate-state dependent friction relation, because of its complexity, are for this model.
The response of the slip weakening model is illustrated by the well-known graphical construction originated by Rice [1979] and depicted in Figure 3. In Figure 3c the solution for the stress and the slip of the slider is given by point B: the intersection of the the slip weakening T versus 6 curve with the line of slope -k, representing the unloading stiffness of the spring. Figure 3d shows the evolution of the system as the load-point displacement 6.
is increased. In Figure 3d the spring is sufficiently stiff (large enough k), that the slider velocity remains bounded as the stress traverses the peak of the T versus 6 curve. Although the rate of slider slip increases, no instability occurs. If, however, the spring is sufficiently compliant (small enough k), then an instability occurs when the slope of the T versus 6 curve becomes equal to -k (point I in Figure 3e). At this point, the slip rate of

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the block 5 becomes unbounded for a finite load-point displacement rate, i.e. finite %.. Thus, instability results from the interaction of the slip weakening constitutive relation with the unloading stiffness of the spring. Figure 3f illustrates a general precursory feature of models based on this constitutive relation: the ratio of the increments of block slip 6 R to imposed slip 6, becomes larger as the instability point is approached.
The effects of certain types of time dependent behavior on precursory processes can also be illustrated in terms of this graphical construction. Figure 4a shows the delay in instability that occurs when the short-time response of the material surrounding the fault is stiffen than the long-time response. As the instability point, predicted on the basis of the long-time or relaxed response, is approached (point I in Figure 4a), the acceleration of slip on the fault induces the stiffen shont-time nesponse. Thus, instability is delayed fnom I to I' in Figune 4a. This type of time-dependent nesponse can anise fnom mechanical coupling of defonmation to pone fluid diffusion [Rice and Rudnicki, 1979; Rudnicki, 1979], fnom stness connosion effects in the matenial sunnounding the fault on fnom coupling of an elastic lithosphene to a viscoelastic substnate [Rice, 1980; Lehnen, Li, and Rice, 1981; Li and Rice, 1983a,b], Of counse, the time scales associated with these diffenent mechanisms ane diffenent.
Time dependence may also enten via the nesponse of the fault zone
matenial. Figune 4b illustnates the stabilizing effect of dilatant handening: if dilatancy (Figune Id) accompanying slip occuns mone napidly than fluid mass can diffuse into the newly cneated void space the local pone pnessune decneases incneasing the effective cornpnessive stness (total stness minus pone pnessune) on the slip sunface; this incnease inhibits funthen fnictional sliding. Thus the condition fon instability is not met until the unloading stiffness becomes equal to the slope of the handened nesponse cunve and instability is delayed fnom I to I' in Figune 4b. Rudnicki and Chen [1986] have necently studied in detail dilatant handening accompanying slip fon a simple one-dimensional model intended to simulate expeniments by Mantin [1980], They find that veny small amounts of dilatancy (only a few pencent of the slip) localized on the slip sunface can stabilize slip as long as the ambient pnessune is high enough. If the ambient pnessune is too low, the pnessune neduction caused by dilatancy accompanying napid slip can cause vapon

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to form or gases to come out of solution, reducing the pore fluid bulk modulus. This eliminates the stabilizing effect and causes an abrupt onset of instability.
Instability associated with the state/rate dependent constitutive
description also results from interaction with the spring stiffness, but the possibilities are more complex and varied than for the rate-independent relation. Instability occurs if a perturbation of the sliding velocity from a steady-state value tends to grow so that the net sliding velocity becomes unbounded. Rice and Ruina [1983] have shown that for a wide class of this type of constitutive relations including (1), the response to small perturbations is unstable only if di /dV < 0 and, in addition, the spring

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stiffness is less than a critical value k . Thus, as for the rate-

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independent slip weakening model, instability can be suppressed by a sufficiently stiff system. However, in contrast to the slip weakening model, this conclusion applies only for small perturbations: Gu et al. [1984] have shown the one state variable relation can be unstable to large perturbations even when k > k . For the two-state variable model an even richer range of cr behavior is possible [Gu et al., 1984],
As already mentioned laboratory results at room temperature generally indicate di /dV < 0 and thus steady-state sliding is potentially unstable. However, laboratory results at higher temperature indicating di /dV > 0

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suggest a transition to stable behavior. The implications of this transition for the depth cut-off of seismicity, as explored by Tse and Rice [1986] will be discussed later.

ELASTIC BRITTLE CRACK MODEL

The single degree-of-freedom, spring-slider system is most appropriate
for modelling laboratory experiments in which slip which occurs essentially simultaneously on the entire surface. But for the rate-independent slip weakening model, the limit in which the slip distribution is strongly non-uniform is also easily understood. In this limit, most of the fault surface is either unslipped or has slipped an amount that exceeds 5 and hence is sliding at the residual value of shear stress T p. The transition from unslipped to slipped portions occurs over a distance w behind the fault edge

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in which the shear stress is reduced from the peak value T to T (Figure 5). Advance of the slipping region is assumed to occur when the relative slip at the end of this transition or so-called cohesive zone achieves some critical magnitude (e.g., Rudnicki, [1980]). If the length w of the transition zone is small compared to other relevant lengths in the problem, then this model is equivalent to an elastic brittle shear crack in which the stress ahead of the fault edge is singular as (distance)" 1 / 2 and the slipping portion of the fault sustains the residual value of the shear stress everywhere. The area under the shear stress versus slip curve in excess of the residual stress (Figure 5c) can now be identified with the critical energy release rate G [Palmer and

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Rice, 1973; Rice, 1980; Rudnicki 1980], Advance of the fault is assumed to occur when the energy that would be released by a unit advance of the fault edge equals this critical value

G = G c (4)

where the left hand side is calculated from the solution for the cracked elastic body. For constant G advance will be unstable if G increases with

\+ . advance, but, in general, G will vary with position reflecting the inhomo-

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geneous strength of the fault. For a fault in a three-dimensional body (two-dimensional fault surface) G will also vary with position around the periphery of the fault edge. Although this variation is known for simple shapes such as circles or ellipses in infinite bodies, the effects of perturbations in the slip front and interactions with the free surface have not been explored. Some recent work has been done [Rice, 1985; Gao and Rice, 1986a,b] on the stability of infinitesimal harmonic variations in edges of semi-infinite or circular cracks in infinite bodies, but more work is needed in this area. Critical values of the energy release rate estimated form field studies
are of the order of 10 6 - 10 7 J/m 2 [Ida, 1973; Rudnicki, 1980]. These values

3 are about 10 larger than those inferred from laboratory experiments [Wong, 1982; Rice 1980]. As remarked by Rice [1983], this discrepancy may result because slip in situ does not occur on a single surface, but on a series of surfaces that are offset. Consequently, the larger values of G reflect the

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significant amounts of inelastic deformation that are required to accomodate slip.

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As noted by Rice [1983], there is not yet a corresponding model for the rate/state dependent friction relation. Presumably, such a model would yield a velocity dependent critical fracture energy. In this case, the conditions for stability of advance will also be more complicated than for the rate-independent model.
The applicability of the elastic-brittle crack model depends on the
endzone size u). Because u scales with 6 , the critical sliding distance (Figure la), uncertainties in the appropriate values of 6 in situ lead to a similar uncertainty in u>. Intermediate cases in which the slip is neither so uniform that the single degree-of-freedom model is appropriate nor so non-uniform that the elastic brittle crack model is appropriate are more complex and must be investigated numerically. The constitutive relation for the fault surface must be combined with modelling of the deformation of the surrounding material. When inelastic deformation is confined to a narrow neighborhood of the fault surface, the deformation of the surrounding material can be sensibly idealized as linear elastic. In this case, the shear stress T at a point P of the fault surface must be related to the slip (assumed to be unidirectional) 6(P',t) at points P 1 as follows (e.g. Rice [1983]):

x(P,t) = T (P,t) - / K(P,P')6(P',t)dS(P') (5)
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where T (P,t) is the shear stress that would be induced on the fault surface in the absence of slip and K(P,P') is the elastostatic Green's function. The stress T (P,t) is assumed to be known and increasing at a slow rate due to large scale plate motions. If the surrounding material is modelled as linearly viscoelastic or the elastic crust is coupled to a viscoelastic asthenosphere, the Green's function will depend additionally on time and a further integration over past history is required. Implicit in (5) is the assumption that the fault surface is straight so that changes in normal and shear stress are not coupled via the slip.
As noted by Rice [1983], when (5) is used with either slip weakening or the rate/state dependent constitutive relations, instabilities analogous to those described for the single degree-of-freedom system can occur. More specifically, d6/dt may become unbounded for some set of points for which di /dt is finite, and this is interpretted as an earthquake instability. In

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the next section, some applications of this type of modelling to crustal faults are reviewed.

APPLICATIONS TO CRUSTAL FAULTING

A complete description of slip progression on a planar surface in
response to farfield loads employing either the rate-independent or rate/state dependent constitutive relations discussed earlier and, possibly, including coupling of the elastic crust to the viscoelastic asthenosphere is a formi-dable three-dimensional nonlinear problem. Consequently, most modelling efforts have considered strike-slip geometries and focussed on particular aspects of the problem, for example, two-dimensional models of slip prog-ression from depth or along strike.

Seminal work in studying the progression of slip from depth on a long (infinite) strike-slip fault was done by Stuart [1979a] and Stuart and Mavko [1979], They used a slip weakening model in which shear stress T at any point on the fault was related to the slip 5 by

T = T p exp {- (6 - <5 Q) 2 /a2} (6)

where 6 is the slip at peak stress T and a is constant that measures the width of the peak. The peak stresses were assumed to have a Gaussian distribution with depth z:

T p (z) = T^ax exp[- (z - z 0) 2 /b2] (7)

where the maximum peak stress t m x occurs at a depth z and the b is a

r p r o constant governing the width of the peak. This variaton is consistent with conceptual models (e.g., Sibson [1982, 1983]) in which strength initially increases with depth due to increasing pressure but then reaches a maximum and decreases due to increasing temperature. This distribution is also consistent with the supposed rr behavior of T p - T p discussed in relation to the slipr weakening model. However, the detailed form of (7) is chosen primarily for convenience. Depending on the parameters entering their model, Stuart [1979a]

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and Stuart and Mavko [1979] find that the response can be stable or unstable and that unstable response is preceded by a rapid straining of the ground sur-face compared with that due to the background rate of tectonic loading. These effects are similar to those discussed for the one degree-of-freedom model. However, an additional feature that does not appear in that model is the progression of slip from depth as points closer to the surface are driven over the peak of the T versus 6 relation. Stuart [1979b] has also used a similar two dimensional model, modified for a thrust fault geometry to study deformation preceding the San Fernando earthquake. He finds that the predictions of the model can be made consistent with uplift observed before and after the earthquake.
Li and Rice [1983a,b] extend the model of Stuart [1979a] and Stuart and Mavko [1979] by including coupling of the elastic crust to a viscoelastic asthenosphere and the effects of a finite rupture length along the strike of the fault. The modelling of Li and Rice [1983a,b] is based upon the "line spring" approximation originated by Rice and Levy [1972] for the analysis of part-through surface cracks. As illustrated in Figure 6, this procedure approximates solution of the three dimensional problem (Figure 6a) by two compatible two-dimensional problems (Figure 6b,c). Rupture propagation along strike is described by a relation between stress and slip averaged through the thickness of a thin plate (Figure 6b) overlying a viscoelastic substrate. This is the generalized Elsasser model [Elsasser, 1969; Rice, 1980; Lehner et al., 1981]. At each position along strike the relation between thickness averaged stress and slip is obtained by solving the anti-plane problem shown in Figure 6c.
In employing the line spring procedure with the elastic brittle crack idealization, Li and Rice [1983a] use a Gaussian distribution of the critical energy release rate to model the upward progression of slip. The maximum of the distribution is assumed to be 7 to 10 km. below the surface, coinciding approximately with the hypocenters of large earthquakes. They also assume conditions are uniform along strike and, as a consequence, the model reduces to a one degree-of-freedom system for which the response is governed by the relation of a single curve of thickness-averaged stress versus thickness averaged slip to the effective unloading stiffness of the surroundings. This stiffness is time-dependent because of coupling to the asthenosphere. Li and

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Rice [1983a] find that this coupling is an important factor in modelling the response immediately prior to instability. The stiffen short-time response delays the onset of instability and gives rise to a period of self-driven precursory deformation as discussed earlier in relation to Figure 4a.
Li and Rice [1983b] discuss the implications of their results [Li and
Rice, 1983a] for precursory surface deformation preceding earthquakes. They predict that the shear strain increases rapidly near the fault but decreases at moderate distances from the plate boundary. The decrease, also noted by Stuart [1981], results from a decrease in the thickness averaged stress. But near the fault this decrease in strain is outweighed by a larger increase due to the approach of upward propagating slip to the ground surface. Unfortunately, the magnitude of the strain decrease is so small that it is unlikely to be detected by current geodetic methods. However, the shear strain increase near the fault is an order of magnitude larger than current geodetic detection levels and the shear strain rate exceeds twice the background rate for time periods ranging from a few months to 5 years, depending on the tectonic loading rate. Li and Rice [1983a,b] also note that the details of their results are dependent on the modelling of the stress versus slip relation in the vicinity of peak stress, but are not sensitive to other features of the constitutive relation. This comment also applies to the work of Stuart [1979a,b] and Stuart and Mavko [1979].
The work just described considers the upward progression of slip in a vertical section, but does not consider variations along strike. Hence, these models are appropriate for analyzing the response on a segment of a long strike-slip fault. Tse et al. [1985] use the line spring approximation to consider effects of variations along strike but they neglect the coupling to the asthenosphere, and do not consider the upward progression of slip. In one calculation, they consider the effects of perturbations in the depth of slip. They find that positions along strike where the slip front is deeper (shallower) the energy release rate (left hand side of (4)) is higher (lower). This suggests a tendency for the deeper portions of the slip front to catch-up to adjacent shallower portions, possibly resulting in minor seismicity. In another calculation, they examine the loading of slip resistant, locked patches by adjacent creeping sections of the fault. As expected, they find that stresses and energy release rates are highest at the edges of the locked

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patches.
Tse et al. [1985] also use this model with slip rate data for the
creeping portion of the San Andreas fault and seismicity patterns near Parkfield to infer the nature of an effectively locked patch near Parkfield. Although, as they note, the inherent inaccuracy of the line spring procedure for describing short wavelength variations prevents detailed resolution of the patch geometry, their results are consistent with work of Stuart et al. [1985], to be discussed below. Tse et al. [1985] also infer a remote stressing rate of 0.3 x 10" 6 u/year where y is the shear modulus and a lithospheric thickness of 30 to 40 km.
Stuart et al. [1985] also study the geometry of the slip resistant patch near Parkfield by means of a three-dimensional model and use their results to forecast the occurrence of the next Parkfield earthquake. They idealize the San Andreas fault near Parkfield as a vertical fault in an elastic slab. The slab extends to a depth of 54 km., chosen somewhat arbitrarily, below which no slip occurs. They assume that each point on the fault is either freely slipping or locked except for a circular patch near Parkfield. The stress and slip on this patch are related by the slip weakening law (6) and the peak stresses are assumed to have a radial Gaussian distribution. The parameters of the model are adjusted so that the predicted surface deformation and creep are consistent with measured values. The model predicts a departure of these data from their background rates as the time of the next Parkfield earthquake approaches. The predicted departure is most notable in the creep records at stations nearest to the inferred location of the resistant patch. Predicted deviations from the background rate are less obvious in the length changes of trilateration lines; the deviation is, however, clearly visible in a line passing just northwest of the patch location, but the modelling of present data is poorest for this line (see Figure 5 of Stuart et al. [1985]). Extrapolation of the results is unable to improve prediction of the time of the next earthquake based on the recurrence interval. Refinement of the prediction may, in principle, be possible with time. The creep and trilateration data cited in the paper extend only to 1982 and no departure of the strain and creep changes from steady accumulation was evident. However, the simple model used by Stuart et al. [1986] appears to be inadequate to fit current data; evidently, it would be necessary to include the interaction

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effects of slip on other nearby faults [W.D. Stuart, personal communication, 1986].
Because of the complexities of the rate/state dependent constitutive relation results have been limited primarily to single degree-of-freedom, spring-slider models. In addition to the frequently referenced but unpub-lished work of Mavko, (see Mavko [1980] for a summary) and the work of Stuart described in this volume, I am aware of only two calculations employing the rate/state dependent constitutive relation for faults embedded in an elastic continuum, those of Tse and Rice [1986] and of Horowitz and Ruina [1986]. Both of these calculations are two dimensional and use the one state variable version of the constitutive relation.
Following the earlier unpublished work of Mavko, Tse and Rice [1986] use
the rate/state dependent constitutive relation on a long vertical strike-slip fault. Although Mavko assumed a plausible depth dependence for the parameters of the constitutive relation, Tse and Rice [1986] infer this depth dependence from observed temperature dependence of laboratory specimens. Laboratory data at room temperature indicates that di /dV < 0. Although there is apparently

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no data between room temperature and 300°C, data from Stesky [1975, 1978] at higher temperatures indicate di /dV > 0. Tse and Rice [1986] convert this temperature dependence to depth dependence by using the Lachenbruch and Sass [1973] geotherm for the San Andreas with the result that di /dV is negative

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(velocity weakening) above 11 km. and is positive below. As discussed earlier instability is possible when di /dV < 0, but not when d-r /dV > 0. Thus, the

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inferred depth dependence provides an explanation of the confinement of earth-quakes to shallow depths. As Tse and Rice [1986] discuss, this explanation is a refinement of conceptual models (e.g. Sibson [1982]) in which the depth cut-off is explained in terms of a transition from brittle to ductile behavior, but suggests that this transition can be described by the depth dependence of constitutive parameters. A noteworthy feature of this explanation is that it does not require a transition from deformation that is localized on the fault at shallow depths to deformation that is more broadly distributed below.
Tse and Rice [1986] use their inferred depth variation on a long vertical strike-slip fault in an elastic plate. Conditions are assumed to be uniform along the strike, but the length enters via the line spring approximation.

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The plate is subjected to a stress increase compatible with a uniform farfield plate velocity. The resulting fault motion is remarkably evocative of many features observed for the occurrence of large earthquakes on the San Andreas fault system. Deep portions of the fault slip continuously at the plate velocity, but shallow portions are effectively locked during most of the earthquake cycles. Driven by continued slip at depth, the slipping zone gradually penetrates upward (as in the previously discussed slip weakening models). Eventually a small region in the locked zone unloads and the spread of this unloading zone initiates an instability. The depth at which instability initiates ranges unpredictably from 3 to 8 km. The instability is followed by rapid afterslip with depth. The cycle then repeats itself with periods ranging from 50 to 160 years for parameters chosen by Tse and Rice [1986].
The most uncertain feature of the calculation is the values of the
parameter L in the constitutive relation (3) to which the results are sensitive. For values of L larger than 160 mm., the entire fault slips stably at the plate velocity. For L between 80 and 160 mm., the shallower portions of the fault do not exhibit earthquake-like instabilities but instead keep up with the deep slip by episodic creep events. Most of the simulations of Tse and Rice [1986] use values of L in the range of 10 to 40 mm. These values are 2 to 3 orders of magnitude larger than values inferred from laboratory tests but may be appropriate in reflecting the larger scale irregularities in the field. Tse and Rice [1986] include one simulation for L = 5 mm., but, because the required mesh size decreases with L the prohibitive cost of simulations prevents a full exploration of results for small values of L.
A result of the calculations particularly relevant to earthquake predic-
tion efforts is the rapid surface slip that precedes instability. However, the magnitude of this slip increases with increasing L (and with decreasing Vdi /dV) and, hence, reflects the uncertainties in the magnitude of L. These

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results and their implications for field measurements are discussed in more detail by Tullis [1987].
Horowitz and Ruina [1986] also use a one-state variable model similar to
that used by Tse and Rice [1986], but differing slightly in the way it models high velocity slip. The constitutive relation is used on an infinite planar fault embedded in an elastic slab loaded by a remote constant velocity. In

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contrast to the depth dependence incorporated in the calculations of Tse and Rice [1986], Horowitz and Ruina [1986] assume that each point on the fault behaves identically. Nonuniformity enters only through the initial conditions. The resulting solutions exhibit a remarkable variety of responses depending on parameter values. Solutions may be periodic, quasi-periodic or aperiodic in time; spatial variations include the possibility of simultaneous occurrence of local creep events, rapid (seismic) slip and steady sliding. The variety of solutions illustrates the richness of fault models based on the rate/state dependent friction relations, but at the same time demonstrates the limited understanding of the models employing these relations for slip between elastic continua. The calculation also suggests at least the possibility that such models may be capable of simulating features of earthquake occurrence that are often attributed to random heterogeneities, e.g., the frequency distribution of different size events.
Although the results of these simulations using rate/state dependent
friction are intriguing and encouraging, the computational resources required for investigation are discouraging. Both the Tse and Rice [1986] and Horowitz and Ruina [1986] models are two dimensional and, although complex in comparison to efforts based on rate-independent constitutive relations, are simplistic in terms of the geometric complexity posed by real faults. In both studies the full exploration of geophysically relevant parameter ranges was inhibited by computational resources (even though the Tse and Rice [1986] calculations were done on a supercomputer). Furthermore, both calculations used the one-state variable constitutive model although two state variables are needed to describe some experimental results. This is a sobering prospect for future research along these lines and suggests that an essential part of this research will be the development of asymptotic methods that will be helpful in improving computational efficiency.

DISCUSSION AND SUGGESTIONS FOR RESEARCH

Current physical models, examples of which have been discussed in this review, are capable of realistically simulating many aspects of crustal earthquakes. Although these models oversimplify the actual geometry of faults and employ constitutive behavior having many uncertain aspects, they produce

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results that are consistent with (or can be made consistent with by adjusting a few parameters) geodetic data. However, current geodetic data is not precise enough to place many constraints on these models or to distinguish among them. More generally, the interpretation of observations is complicated by the likely presence of a variety of competing effects. The modelling efforts have, quite sensibly, focussed on particular aspects, for example, upward progression of slip or coupling of the crust to the asthenosphere. Yet, all of these effects, as well as those due to coupling between deformation and pore fluid diffusion, are likely to contribute to precursory processes. Thus, although physical models have increased greatly in complex-ity and sophistication, there is still much to accomplish.
Better three dimensional modelling is needed not only to provide better comparison with field observations, but also to investigate geometric effects that are not evident in two-dimensional models. As reviewed here, most fault models have been two-dimensional. Some approximate three-dimensional model-ling has been done using the line spring procedure. The great advantage of this approach is its relative simplicity compared with the fully three-dimensional modelling. However, in most cases in which the line spring approach has been implemented, variations in one of the dimensions has been simplified, for example, by assuming that conditions are uniform along strike. When variations over the entire fault plane are considered, the advantageous simplicity of the line spring method may be eroded. Another limitation of the line spring method is, as noted by Tse et al. [1985], its inherent inaccuracy for short wavelength variations. Often, however, abrupt transitions between slipping and effectively locked portions of the fault are of great interest as, for example, at Parkfield. Some of these drawbacks may be overcome by hybrid approaches, for example, combining the line spring procedure with a boundary element method [Li and Fares, 1986], but recourse to direct numerical methods should be considered as an alternative.
Recent three-dimensional modelling has been done by Stuart [Stuart et al. 1985]. His procedure divides the fault plane into a series of cells in which the displacement is uniform. Hence, the displacement is discontinuous and the stresses at the cell boundaries are unbounded. Further, because of the stress stron singularity introduced by the uniform displacement solution, the average stress in each cell is unbounded. Although this method yields satisfactory

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results, it is unlikely that the details of the stress distribution are accurate. The method would also seem to have difficulty in describing the advance of a sharp slip front as, for example, in the elastic brittle crack 1imit.
Although three-dimensional calculations using the rate/ state dependent friction model do not appear to be feasible at present, there is much that can be done with simpler models, particularly the elastic brittle crack model. For the crack models, efficient and accurate procedures for three-dimensional calculations of fracture progression are available [Lee and Keen, 1986; Lee et al., 1986], Conceptual models of earthquakes, based primarily on seismic observations, attribute much significance to slip resistant portions of the fault plane, usually called barriers or asperities. The procedures just-mentioned could be used to explore the progression of slip fronts through or around slip resistant patches on the fault plane. More generally, there has been little exploration of effects of geomtric perturbations in an advancing slip front, perhaps caused by encountering slip resistant zones. Tse et al, [1985] have made some preliminary calculations for the variations in energy release rate due to along-strike depth variations in a stationary slip front. Also Rice [1985.] and Gao and Rice [1986a] have made progress in this direction with analytical calculations for the stability of infinitesimal harmonic vari-atons in the straight front of a crack in an unbounded body. However, they find that the critical wavelength above which perturbations are unstable is sufficiently large that the idealization used by them of a half-plane crack in an infinite body is unsuitable. Consequently, there is a need for further work in this area, including the effects of the free surface. An important goal is to try and understand conditions that distinguish large and small events.

Current physical models simulate realistically large events, but due to
the prohibitive costs of using very fine grid sizes, it has not been possible to model small scale instabilities that could be interpretted as precursory seismicity. This shortcoming is unfortunate because minor seismicity is often routinely and reliably monitored. At present there is no clear path around this difficulty. Possibilities include use of more efficient computational algorithms, asymptotic multiscaling techniques, or more supercomputer time, but it seems unlikely that large computations involving disparate size scales

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will be feasible in the near future. Some progress may, however, be possible again by using crack models if increases in seismicity can be identified with the increase of stresses due to an advancing slip front. Such models can also be used to predict surface deformation and, thus, can be used to relate sur-face deformation to time and space variations in minor seismicity, an object-ive repeatedly emphasized by Rice [1984, 1985],
Because of the difficulties of including minor instabilities, represen-tative of minor seismicity, in continuum calculations there has been much work on multi degree-of-freedom spring-block systems. (See Cao and Aki [1986] or Nussbaum and Ruina [1986] for summaries.) The conclusions of these studies have been varied and seem difficult to reconcile with one another, suggesting a strong sensitivity to choice of parameters. A persistent question has been the physical basis of observed seismicity patterns and, more specifically, the magnitude frequency relation. To what extent are these attributable to fault surface heterogeneity, fault surface constitutive relation, and/or some as yet unidentified characteristic of a dynamical system model of the earth? Recent simulations by Cao and Aki using slip weakening [Cao and Aki, 1984/85] and rate/state dependent [Cao and Aki, 1986] friction relations have produced reasonably realistic seismicity patterns. The simulations using the slip weakening law were not, however, capable of continually repeatable, production of small earthquakes. Because the rate/state dependent friction relation does include restrengthening of fault segments that are sliding at low velocities, a heterogeneous distribution of strength can be continually regenerated on the fault. Consequently, on the basis of their simulations using the rate/state dependent friction relation, Cao and Aki [1986] conclude that this relation provides a physical mechanism for the roughening process that is necessary to explain the observed stationary magnitude - frequency relation. However, Nussbaum and Ruina [1986] have shown that a simple model with two identical degrees-of-freedom and employing a static/dynamic friction almost always tends toward an inhomogeneous state after many slip events. Moreover, one class of solutions is found to be structurally unstable in the sense that the results are strongly sensitive to perburbation. Although the simplicity of their model makes direct application to the earth difficult, their results suggest the possibility that spatial seismicity patterns may be due, at least in part, to nonlinear interactions in a complex dynamical system.

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The further development of physical models is inhibited by uncertainties
in how to extrapolate laboratory measurements to the crust, in particular measurements of the critical distance parameters entering the friction relations. This difficulty is compounded because the appropriate discreti-zation in crustal scale models is tied to the size of this length scale and computational costs increase as the size scale diminishes. This effort would be aided by a better understanding of the micromechanical mechanisms that give rise to the observed phenomenology. Current efforts on scaling have concen-trated on the effects of surface roughness, but variations in fractional behavior among different rock types suggests that the behavior may also be related to constituent minerals or whole rock deformation at pressures approximating those at asperity contacts. A possibility that cannot be

ignored is that slip on faults in the earth is quite different from that on / well-defined surface or gouge layers in the laboratory. Rice [1983] has pointed out that the discrepancy between values of the critical energy inferred from laboratory and field studies may result because macroscopic fault slip in situ involves slip on offset segments requiring considerable inelastic deformation between them. Such processes may also be relevant to the larger values of the critical sliding distance inferred from crustal fault models.
Although models have, for the most part, considered straight faults,
bends are important because the normal stress induced across the fault can inhibit slip. Stuart. [1986] does approximate the curved San Andreas by a series of straight line segments, but evidently does not account for induced variations in normal stress. The effects of normal stress on the slip weakening constitutive relation are reasonably well understood [Wong, 1986], but there is , yet, only preliminary information on the effects of normal stress changes in the rate/state dependent model [Olsson, 1985; Hobbs and Brady, 1985; Linker and Dieterich, 1986],
In all the models discussed here the inelasticity is confined to the
fault surface. This would seem to rule out modelling precursory phenomena associated with widespread cracking or dilatancy, such as resistivity changes, radon transport or travel time changes. Models based on embedded volumes of inelastically deforming material have been suggested [Rudnicki, 1977, Rice and Rudnicki, 1979], However, there has been no mechanism convincingly demon-

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strated for causing inelasticity to be spread from a narrow zone, although dilatancy hardening is a possiblity [Rice, 1975; Rudnicki, 1984; Rudnicki and Chen, 1986]. It may be that widespread inelasticity is induced in the vicinity of bends and offsets in the fault trace. These have been suggested as places where ruptures start and stop [Sibson, 1986] and the mechanics of these regions warrants more study.

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FIGURE CAPTIONS

1. (a) Idealized experiment in which slip <5 is assumed to occur simultaneously everywhere on the fault surface. T is the shear stress, a
is the normal stress, A is the opening and p is the pore pressure.
(b) Slip weakening friction relation. The shear stress increases with

slip 6 to a peak T and then decreases to a residual value T after an r r p r
amount of slip 6 . Sliding continues at T = T for 6 > 6 .
(c) Variation of T and T with effective normal stress a - p.
(d) Dilatancy A accompanying slip.

2. (a) Idealized experiment in which sliding occurs at a steady slip-rate
5 = V.

(b) Change in shear stress after a sudden increase in the sliding velocity from V, to V«.

3. (a) One degree of freedom, spring-slider model for illustrating instability. The imposed displacement is 5,(t), the spring stiffness is
k, the displacement of the block is 6 R > the friction stress exerted on
the block is TD and the constant normal stress is a.
(b) An isolated fault of length 2a, sustaining a uniform friction stress
T^ and loaded by a farfield stress T . f «
(c) Graphical solution for the motion of the system in (a) with the slip weakening constitutive relation. Solution is given by point B.
(d) Evolution of the system as the imposed displacement increases. The figure illustrates a case for which the spring is stiff enough that instability does not occur.
(e) Instability corresponding to dS B /d6, > °° occurs at point I when the stress versus slip curve becomes tangent to the spring unloading line.
(f) Illustrates the acceleration of fault slip preceding instability. Figure shows increasing increments of 6^ corresponding to equal increments of 6.

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4. Delay of instability and enhancement of precursory slip in the slip-weakening model by time-dependent effects.
(a) Elastically stiffen response of the surroundings to rapid straining delays instability from I to I 1. Such time-dependence can arise from coupling of deformation with pore fluid diffusion, coupling of an elastic lithosphere with a viscoelastic asthenosphere or stress corrosion effects
in the material surrounding the fault.
(b) Hardening of fractional response to rapid slip can also delay
instability from I to I 1. Such hardening may arise from reduction of pore pressure (and increase of effective compressive stress) due to dilatancy accompanying rapid slip.
5. Cohesive zone fault model.
(a) Variation of slip: for x > 0, 6 = 0 and for x < w, 6 > 6 .
(b) Corresponding variation in stress. Note T = T for x < -01.
(c) Slip weakening relation between shear stress and slip. When the
cohesive zone size 01 is small compared to the overall fault length and
other dimensions, the shaded area under the T versus 6 curve can be

identified as the critical energy release rate G for a shear fault. L*

6. Illustration of the line-spring approximation.
(a) A long strike-slip fault. The depth of slip below the surface varies along the strike of the fault (x direction). This three dimensional prob-lem is approximated by the two dimensional problems shown in (b) and (c).
(b) Thin plate approximation is used to relate variations of thickness-averaged stress and slip along the strike (x direction). The thickness-averaged relation between stress and slip at each section (each x) is obtained by solution of the anti-plane problem in (c).

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(a)

long-time, relaxed
(drained)

.short-time, unrelaxed Cundrained)

short-time (dilatantly)
hardened

long-time (drained)

(b)

S

Figure 4
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Forecast Model For Great Earthquakes at the Nankai Trough Subduction Zone

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Abstract - An earthquake instability model is formulated for recurring great earth-quakes at the Nankai Trough subduction zone in southwest Japan. The model is qua-sistatic, two-dimensional, and has a displacement and velocity dependent constitutive law applied at the fault plane. A constant rate of fault slip at depth represents forcing due to relative motion of the Philippine Sea and Eurasian plates. The model simulates fault slip and stress for all parts of repeated earthquake cycles, including post-, inter-, pre-, and coseismic stages. Calculated ground uplift is in agreement with most of the main features of elevation changes observed before and after the M=8.1 1946 Nankaido earthquake. In model simulations, accelerating fault slip occurs shortly before unstable (earthquake) fault-ing. The pre-instability accelerating fault slip has two time-scales. The first time-scale is several years long and is interpreted as an intermediate-term precursor. The second time-scale is a few days long and is interpreted as a short-term precursor. Accelerating fault slip on both time-scales causes anomalous elevation changes of the ground surface over the fault plane of 100 mm or less within 50 km of the fault trace.

Introduction

In prior work (e.g., Stuart, 1986; Stuart et a/., 1985; Stuart and Mavko, 1979)
I suggested that earthquake instability models might have application to earthquake prediction because the models simulate faulting and ground deformation for all stages of an earthquake cycle. The prediction strategy would be to use ground motions observed during inter- and preseismic stages to estimate values of model parameters. This step can be accomplished by fitting part of a model simulation to available observations either by trial or inversion. Once parameter values are fixed, the model can be used to calculate the time, location, and amount of subsequent unstable slip (if any) because a specific set of parameter values yields a unique faulting history for all stages. In practice, sparse and imprecise field data and lack of realism in models would limit the resolution of model parameters and thus also the accuracy of an earthquake prediction. Another practical limitation is that a parameter value may be resolvable by data from only one seismic stage since individual model stages are dominated by a subset of all parameters.
This prediction procedure resembles numerical weather forecasting (e.g., Tribbia and Anthes, 1987) in that both methods combine repeated measurements of field data with a well-posed mechanical model in which the phenomena of interest (seismic and aseismic faulting, atmospheric flow) are computed results. At present, instability models appear

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to be the only class of fault models that is strictly applicable to prediction in this sense. Other mechanical models, such as those in which an earthquake is asserted to occur at a critical stress or strain level do not unify pre- and coseismic faulting, do not involve a plausible fault constitutive law, or are not posed as boundary value problems that might be solved. Such models also contradict the well-known result from fracture mechanics that the condition for faulting instability is loss of equilibrium, not critical stress or strain (e.g., Lawn and Wilshaw, 1975, p. 1).
Instability models are especially well-suited to intermediate-term earthquake predic-tion (weeks to years) because they predict accelerating ground deformation on that time-scale. The accelerating fault slip occurs in about the last 5% of the recurrence time before instability (e.g., Stuart, 1986). In contrast, instability models may have less value for long-term prediction unless the models could account for observed variations of recurrence times. The usefulness of instability models for short-term prediction (seconds to weeks) is little studied. As discussed elsewhere (Rice, 1983; Stuart, 1986), published instability models appear to be consistent with the main features of seismic and aseismic faulting for several specific earthquake regions, but the models need more testing. Additional testing is especially desirable for the period of accelerating fault slip and ground deformation that would be the main basis of an intermediate- or short-term prediction attempt.
There have been two applications of instability models to specific prediction attempts, both for sections of the San Andreas fault in California. One application (Stuart et a/., 1985) is to the fault near Parkfield, where a moderate earthquake is expected. The other (Stuart, 1984/85, 1986) is to southern California, where large and great earthquakes are expected. In both cases the models represent the San Andreas fault by a single vertical fault plane.
This paper attempts to determine if instability models could have similar application
to forecasting great earthquakes at subduction zones near Japan. Instead of constructing ab initio a model for the entire region, which is tectonically complicated, I start with a simple model for a portion of the Nankai Trough extending southwest from the Tokai seismic gap. The subduction zone at the Nankai Trough has had recurring great earthquakes at least since 684 AD (Ando, 1975) and thus is expected to have more in the future. The model is tested with faulting and crustal deformation data associated with the most recent great earthquake, the M=8.1 1946 Nankaido event. The two main goals are to see if an instability model is consistent with available data and to determine if predicted ground uplift anomalies leading up to such earthquakes are large enough to detect.
The next section describes the mathematical model and its solution. The following section applies the model to the Nankaido earthquake and compares theoretical and observed uplift. The main results are that the model is generally consistent with available uplift data, and that theoretical preseismic ground uplift is large enough to detect.

Instability model

The model is made of pieces of previously published models plus a few minor innovations. The fault plane, shown in cross section in Fig. 1, extends from the ground surface y = 0 downward in the £ direction and indefinitely normal to the x-y plane. At

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depths above £ = d shear stress and slip velocity on the fault are required to satisfy a constitutive relation to be discussed shortly. Below depth £ = d a constant slip velocity Vp representing relative motion of tectonic plates is imposed. The uniform slip due to Vp is equivalent to a single edge dislocation of increasing Burgers vector at d as shown in Fig. 1. The growing dislocation stress loads the upper part of the fault above £ = d to recurring failures, each failure defining the end of an earthquake cycle.
The fault above £ = d is divided into two parts according to its assumed constitutive behavior. The upper one-third has a predominantly brittle behavior such that the fault shear stress resisting slippage can decrease with increasing slip if the fault has prior strength. In contrast, the lower two-thirds has a viscous-like behavior such that fault stress increases with slip rate. In qualitative terms the justification for the two different kinds of fault behavior is that brittle behavior characterized by slip weakening is required to produce instability, whereas viscous behavior is required to limit the down-dip penetration of unstable faulting and to explain certain postseismic crustal motions. This partitioning of fault behavior according to depth is consistent with laboratory results that show brittle deformation at low temperatures and confining pressures and viscous behavior at high temperatures and pressures.
There are a number of mathematical ways to represent brittle and viscous constitutive laws. The constitutive law used here is from Rice (1983), Rice and Tse (1986), and Tse and Rice (1986) and is motivated by laboratory friction experiments (e.g., Dieterich, 1981). It is most conveniently expressed as the differential equation

dr/dt = (dr/dV)-msi dV/dt - (V/L){r - ra8 (V)} (1)

where r is shear stress, t is time, and V is slip velocity.
The first term on the right side of equation (1) multiplied by dt is interpreted as the shear stress increment caused by an instantaneous jump of sliding velocity. The form

(dr/dV)-msi = A/V (2)

for the first parenthetical term in (1) provides a reasonable fit to laboratory data. Coefficient A is a positive material constant for fixed temperature and effective normal
stress.

The second term on the right side of equation (1) is interpreted as the initial exponential path that shear stress follows as it approaches the new steady state value of T SS (V) that depends on the post-jump velocity; L is the characteristic slip for the decay of r to T sa (V). In an application to instability on crustal scale strike slip faults, Tse and Rice (1986) considered two forms for T SS. The form used here is

T "(V) - r* = -(A - B)]n(V*/V + e- 10) (3)

which allows for rss to level off at velocities greater than about 0.1 mm/sec, in accord with experiment. Stress r* is a reference shear stress such that r 38 = r* in (3) (neglecting e~ 10) when the fault slips at reference velocity V* Vp. The term A B is a material constant for fixed temperature and effective normal stress. When A B is positive it gives

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a viscous-like increase of rss with increasing velocity, and when it is negative it gives a decrease of rss with increasing velocity and the possibility of instability.
In the model below, A and A B vary with position on the fault since temperature and confining pressure increase with depth in the earth. L is assumed to be independent of position on the fault and of V. In view of (3) there are two cases to consider when A B is negative. When the velocity jump is positive, shear stress decreases with increasing slip and is equivalent to slip weakening. When the velocity jump is negative, the law produces increased resistance to slippage, equivalent to a healing of the fault. Thus in a model simulation the lull of slip rate on the brittle part of the fault after instability is self-reinforcing and responsible for subsequent instabilities.
Rice (1983) and Rice and Tse (1986) gave (1) as the equivalent of a general class of single state variable forms for constitutive laws. This class includes the laws developed by Ruina (1983), which in turn were approximations to empirical equations described in Dieterich (1981). Fault law (1) is expressed as a single differential equation with no reference to the state variable, whereas in formulations of Kosloff and Liu (1980) and Ruina (1983) shear stress is expressed as a transcendental equation and the state variable as a differential equation. The single state variable law of Kosloff and Liu (1980), which is the law of Dieterich rewritten to express the state variable as a differential equation, is also representable by (1). The instability model of Stuart (1986) and its predecessors use a simpler constitutive law that is rate independent and thus has no viscous or strengthening capability. The study by Mavko (unpublished manuscript, 1984) of instability on crustal scale strike slip faults used a state variable form by Ruina (1983).
The experimental data motivating equation (1) are special enough that (1) can be obtained in a simple way and without explicit use of a state variable. The experimental data used in the construction of (1) are from a single kind of ideal test, namely a step change in velocity from one value to a new and constant value. Associated with the instantaneous velocity jump is an instantaneous shear stress jump, but no change of the state variable or slip. After the velocity jump, shear stress is assumed to exponentially approach a steady state shear stress rss (V).
First, note that the single state variable equivalent of (1) studied by Rice (1983) and Rice and Tse (1986) starts with the postulated functional forms T = F(V, 0) and dO/dt = G(V, 0), where 0 is the state variable, and temperature and normal stress are here constant. The state variable itself has not yet been associated with a measureable physical quantity and thus is not accessible by experiment or study of field data. Without further use of G(V, 0), the total differential of F(V, 0) and the above experimental results are sufficient to constrain the forms of terms in (1). The total differential of F(V, 0) divided by dt is

dr/dt = (drldV)e dV/dt + (dr/d0) v dO/dt (4)

Subscripts 0 and V indicate that those quantities are held constant during the differentia-tion. The first term on the right side is the same as in (1) and represents the instantaneous jump of stress at constant state. The second term on the right side is just the stress rate at fixed velocity immediately after the velocity jump. If, as Rice (1983) assumes, the variation of stress with slip satisfies an exponential decay, then dr/dU = (r rss)/L, where U is

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slip measured with respect to slip at the time of the velocity jump. Substituting dU = Vdt gives dr/dt at constant V, the same as the last term in (1).
Now, since equations (1) and (4) are constrained only by experimental observations of
r, U, and V at and following a velocity jump, it seems worthwhile to obtain dr/dt directly in terms of those variables. For a test where velocity is a step function assume that shear stress T = r(C7, V;i). Then the total derivative is

dr/dt = (dT/dV)u dV/dt + (dr/dU) v dU/dt (5)

As above, subscripts denote quantities held constant during differentiation. The first right side term represents the instantaneous stress change at constant slip (constant state), as in (1). Substituting into the last term of (5) the expressions (dr/dU)v = (T rss)/L and V = dU/dt gives the last term in (1).
The objection to assuming r = r(C7, V) for a general observed stress history which might contain repeated instabilities is that there can be no fundamental dependence of stress on slip. However, r r(U,V) is certainly valid for a single velocity jump test, and in numerical simulations of a stress history a smooth velocity curve is approximated by a staircase function having sufficiently small AF and A£.
The remaining constraint on the model is that static equilibrium holds at all points
on the fault plane, that is, the stress exerted by the deformed elastic earth balances the resisting stress from the fault law. This condition can be expressed as an integral equation involving the forcing due to Vp, the distribution of dislocations on the fault, and the fault law as in Stuart and Mavko (1979), but the equation is too difficult to solve analytically. To obtain a numerical solution for T and V as functions of £, the equilibrium condition is replaced by a set of nonlinear simultaneous equations which result from dividing the fault into n equal segments of uniform slip, as shown in Fig. 2. The elastic stress from the earth acting on a small section of the fault is approximated by the stress at the center of each segment i and is given by

Kij (Vj - Vp) (6)
j = i

The variable Kij, an elastic stiffness, is the shear stress due to unit slip on segment j acting at the center of segment i. The elements of matrix K are calculated from an analytic solution in Freund and Barnett (1976, eqn. 13) (see also Dmowska and Kostrov, 1973). (The errata, Freund and Barnett (1976), has misprints. The denominators of the last terms in equations (4) and (5) should read (z ~z0) 2 and B(z -\-~z 0) respectively.) Equation (6) is equivalent to the method of Savage (1983) for dividing fault slip into steady and time-dependent components, and thus the stress field from the single dislocation representing Vp in Fig. 1 need not be considered explicitly. Combining equations (1) and (6) so as to satisfy equilibrium yields the system of simultaneous first order differential equations (7).

n

dn/dt = - KH (Vi - vp)
J = l

dVi/dt = -(VIA) dTi/dt + (V 2 /AL) [T - rss (V)}

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The reference stress r* does not appear, and only variations of stress, not absolute levels of stress, are involved with instability (cf. Tse and Rice, 1986).
The model does not allow for fluctuations of normal stress on the fault caused by fault slip. Such normal stresses occur for dip-slip on a flat fault plane when a free surface is present. Additional normal stresses would also occur if the fault plane had curvature or bends. Both kinds of normal stresses can be calculated from solutions in Freund and Barnett (1976) and used to produce an additional set of equations like (6). Equations (7) would then be replaced by a set of equations for shear stresses, a set for normal stresses, and a set containing the fault law modified to include normal stress changes.
Equations (7) are subject to initial conditions approximating the velocity and stress state just after unstable faulting. A satisfactory set of initial conditions is V = V\\m, r = T ss (V\\m) on the interval from £ = 0 to the depth where A B changes from negative to positive. The variable Vjim is the maximum velocity allowed in the model and is set equal to 1 m/sec, a nominal slip velocity during seismic rupture. Between the position where A B changes sign and depth d, V decreases linearly to Vp, and r equals rss (V). As Tse and Rice (1986) note, when such an initial condition is used, successive faulting cycles seem to approach a limit cycle in which the immediate post-instability state is similar to the initial conditions just described.
The set of differential equations (7) is integrated forward in time with a fifth order Runge-Kutta procedure modified from one in Press et al. (1986, p. 550). At each time step, U is calculated from integration of prior velocity. The time step size is automatically adjusted during the Runge-Kutta integration so that the fractional truncation error for every fault segment is less than 10~ 4. Instability is defined to occur when the velocity of at least one fault segment exceeds V\\m. When this condition is met, the unstable slip is estimated by a separate iterative procedure described in Tse and Rice (1986). In the procedure, the upper and lower ends of unstable faulting are those at which further extensions of the ends of a trial fault length first give a stress increase ahead of the shear crack tips instead of a decrease.

Application to the 1946 Nankaido earthquake

The application of the instability model to the 1946 Nankaido earthquake is presented
in three steps. The first step is to find values of model parameters that give the recurrence time between the 1854 Ansei earthquakes and the Nankaido earthquake, and also agreement between simulated and observed coseismic uplift for the Nankaido earthquake. The free parameters are L and the value and position of the most negative A B, called (A -B) min. This simulation and its set of parameter values is referred to as the reference case. The second step, a check of the model fit, is a comparison of theoretical and observed uplift profiles for other stages of the earthquake cycle. The last step is a search for intermediate-and short-term precursors in theoretical uplift curves.

Reference case

Fig. 3 shows the location of the Nankai Trough subduction zone and the direction of relative motion between the Philippine Sea plate and the Eurasian plate. The rectangles are projections to the ground surface of coseismic dislocations for the 1944 Tonankai (M=8.1)

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and 1946 Nankaido earthquakes inferred by Ando (1975) from analysis of horizontal and vertical geodetic data. Ando represented the Nankaido earthquake by two dislocations with 6 m and 4 m slips as shown. The respective dips of the fault planes are 20° and 25°. The slip and dip for the Tonankai dislocation are 4 m and 25°. Dots with vectors show epicenters and seismic slip directions determined by Kanamori (1972) from seismic records.
Ando (1975) and Mogi (I985a) summarize evidence that the Nankai Trough plate boundary repeatedly ruptures by great earthquakes. The earliest such earthquake mentioned by Ando (1975) was in 684 AD, but the earliest well-documented event was the 1707 Hoei earthquake. Its rupture surface extended from the southeast end of the Nankaido rupture to about 100 km beyond the northwest end of the Tonankai rupture (Fig. 3). The 1854 Ansei I and II earthquakes (32 hours apart) together covered the same area, with the Ansei II earthquake having roughly the same slip surface as the Nankaido earthquake. The Tonankai earthquake ruptured about half the plate boundary length that the Ansei I earthquake did; the unruptured remainder is the current Tokai seismic gap (e.g., Mogi, 1985a).
I will apply the instability model to the part of the subduction zone under the Kii Peninsula, as shown by the model x axis in Fig. 3. Thus the model will be most appropriate to the narrow half of the Nankaido rupture inferred by Ando (1975) and will be mainly constrained by levelling data measured along the route from Osaka to Kushimoto (dashed line). Features of the seismic cycle revealed by other level lines and tide gages shown in Fig. 3 are similar and are described by Thatcher (1984) and Thatcher and Rundle (1984) but are unused here. A more complete treatment of the data would require a three-dimensional instability model.
Since the instability model produces repeated earthquake cycles, with faulting during successive cycles generally growing more alike with time, a particular instability must be chosen to represent the Nankaido earthquake. The first model cycle is defined to start with initial conditions which approximate an immediate post-instability state, and end immediately after the first instability. Faulting during the second cycle is essentially the same as during subsequent cycles and therefore the second instability is taken to be the analog of the Nankaido earthquake. The second instability will be identified on Fig. 5 shortly.
The heavy dashed and solid lines in Fig. 4 show the down-dip variation of A (equation
2) and A B (equation 3) for the reference case. The curves have nearly the same shapes as in Tse and Rice (1986), but distances are scaled to the Nankaido earthquake. The amplitudes of A and A B end-points and slope changes are the same as in Tse and Rice (1986) except for (A B)m-m. Both A and A B have rate increases at f = 120 km. However, a simulation with no slope change in A and A B as shown by the light line in Fig. 4 differs from the reference case by less than a few percent. In the antiplane model of Tse and Rice (1986) the physical reason for such slope increases of A and A B is that temperature replaces confining pressure as the main control on constitutive behavior. Tse and Rice (1986) use a profile of crustal temperature versus depth estimated from heat flow measurements made near the San Andreas fault, but here geodetic data are thought to be more reliable constraints on constitutive law parameters than are estimates

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of temperature at depth. The parameter A jE?, which is the dependence of steady state stress on velocity, has sections of negative and positive value. As discussed above, the physical interpretation of negative and positive intervals of A B is, loosely speaking, brittle and ductile behavior of the fault. In simulations the amount of unstable slip and stress drop increases as (.A B) m-m becomes more negative. No instability occurs in simulations if A B is non-negative everywhere, only uniform velocity with depth after initial transients decay.
Tse and Rice (1986) used actual or interpolated laboratory values of A and A B measured at conditions corresponding to estimates of confining pressure and temperature in the earth. This approach seems to have been successful, but geodetic data were unavailable to check the resolution of A, A B, and L. Although laboratory values of A and A B were satisfactory, Tse and Rice (1986) concluded that laboratory values of L are about 10~3 too small if L is interpreted as a critical slip weakening distance for crustal scale earthquakes. Tse and Rice (1986) found that values of L between 10 and 40 mm resulted in simulated strike slip earthquakes of several meters slip, but limits on computer time precluded study of much smaller L. The range of permissible L in the Nankaido model is discussed below. It is unclear why some laboratory values seem appropriate to the earth and others do not. On the other hand, Lockner and Byerlee (1986) point out that laboratory estimates of A B calculated from results reported by various workers using different experimental configurations and load programs have wide scatter. In particular they note that A B, which must be negative for instability to occur with equation (l) under steady loading, is often positive in laboratory experiments. Some of the laboratory experiments reported in Lockner and Byerlee (1986, Figs. 10 and 13) as giving positive A B are associated with stick slip or oscillatory slippage. The occurrence of stick slip, of course, requires that the testing machine and intact portions of the rock sample be sufficiently compliant.
The free parameters found by trial for the reference case are (.A B) m\ Q = 0.8 bar
at £ = 55. km, and L = 40 mm. Other parameters have assumed values and are crustal rigidity of 0.3 mbar, Vp = 40 mm/yr (Seno, 1977), and fault dip a = 25°. The actual dip of the subducting Pacific plate in the vicinity of the model x axis in Fig. 3 is variable. Mizoue et al. (1983) and Yamazaki and Ooida (1985) show that microseismicity at the top of the plate defines a plane dipping at about 10° between the Nankai Trough and the northwest edge of the Ando dislocation. From the dislocation edge to about 40 km farther northwest, dip increases to about 35°. The 25° dip of the model fault gives the best overall fit to measured uplift of any single flat plane.
In the numerical solution one hundred fault segments (Fig. 2) are used, each 1.5 km long. The minimum time step size needed to maintain numerical accuracy is 10~ n yr (0.003 sec), and it is used just before and after instabilities. The maximum time step size is about 2 years, occurring midway in time between instabilities.
Fig. 5a is a plot of fault slip versus down-dip distance at successive times for the reference case. The time interval between lines is not constant because the time step size in the simulation automatically decreases when the slip rate at any point increases. At the right side of Fig. 5a, fault slip increases at a nearly constant rate, whereas at the left side, slip rate alternates between high and low values. High slip rate occurs during pre-, co-, and postseismic faulting, and low rates when the fault is essentially locked during the

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interseismic stage. On the plot the fault is locked where the lines on the left are nearly horizontal above £ = 60 km and most closely spaced. Instabilities representing the 1854 Ansei II and the 1946 Nankaido earthquakes are marked and appear as jumps in slip. Fig. 5b shows more clearly the slip changes during stages of the earthquake cycle containing the Nankaido earthquake. Each stage is bounded by two lines, and each line is labelled with a time measured with respect to the instant of instability 2. Lines in Fig. 5b are either the same as in Fig. 5a or between two lines in Fig. 5a. Using earthquake terminology, the model stages are labelled interseismic, intermediate-term precursor, short-term precursor, coseismic, short-term postseismic, and intermediate-term postseismic. The precursory stages involve accelerating fault slip along the upper 60 km of the fault and have time-scales of a few years and a few days respectively. The postseismic stages involve decelerating and downward propagating slip along the upper 100 km and have time-scales of about 4 days and a year respectively. Unstable slip occurs in zero time since the model is quasistatic.
Fault slip during each stage of Fig. 5b is for the most part either on the brittle
(£ < 58 km) or on the viscous (£ > 58 km) section of the fault. Preseismic faulting is confined to the brittle fault because it is driven by elastic energy stored during the long interseismic stage. Unstable slip is similarly caused by failure of the brittle fault, but its downward penetration is limited by the viscous fault. The intermediate-term postseismic faulting is driven mainly by the stress concentration left near 70 km by the preceding unstable slip.
Figs. 6a-6d show qualitatively how the amount of pre- and postseismic fault slip relative to coseismic slip decreases as the intensity of instability increases. A highly unstable case has more unstable slip compared to pre- and post-instability slip on the brittle fault than does a slightly unstable case. Panels (a)-(d) of Fig. 6 are in order of increasing instability and correspond to different pairs of (A B) mm - and L. Each panel shows fault slip versus down-dip distance at successive times like in Fig. 5a except that four cycles are shown here. Fig. 6b, the reference case, includes Fig. 5a. All cases have instability recurrence times between 88 and 95 years. The overall amount of preseismic and postseismic slip on £ < 60 km clearly decreases going from (a) to (d), while at the same time coseismic slip increases, the sum being nearly invariant. Usually the form of preseismic slip has a modest variation from one instability to the next, but exceptions occur in (b) and (c). Recalling from Fig. 5b the form of slip during the intermediate-term precursor stage, it is clear that the amplitudes decrease from Figs. 6a to 6d, though the shapes are similar. Short-term precursory slip, as indicated by the stack of overlying lines, undergoes a much greater shape change and amplitude decrease going from (a) to (d). Simulation (b) in Fig. 6 is chosen to be the reference case over others because it has both substantial precursory slip and unstable slip. This compromise is suggested by the average seismic slip of 3 m for the Nankaido earthquake (Kanamori, 1972), the 4 m coseismic slip estimated by Ando (1975) from geodetic measurements made several months after the earthquake, and possible short-term uplift precursors (Sato, 1977; Mogi, 1985b). Fig. 7 summarizes the results of using different values of (A B) m-m and L while holding other parameter values fixed. Each dot corresponds to a simulation, and the dots marked 'a'-'d' refer to the simulations in Fig. 6. The circled dot 'b' is the reference

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case. Contours show the recurrence time between the first and second instabilities for each case, and thus cases suited to the Nankaido earthquake must lie near a 92 year contour. Contour lines also define the approximate direction of increasing intensity of instability, which is from the upper right to the lower left. A stable case may have repeated episodes of high slip rate but not any unstable slippage. Since no instabilities occur in the region marked 'stable', no cases there can represent the Nankaido earthquake. On the other hand, highly unstable cases at the lower left of the plot caused by very negative (A JE?) mi n and small L give elevation changes during instability that are large compared to pre-instability changes. Thus a few observations restrict the domain of acceptable cases in ((A J5) min, L) parameter space; the difference between cases within the domain is mainly the degree of instability, which cannot be distinguished here due to scarce geodetic data just before and after the Nankaido earthquake.
Highly unstable cases also tend to have short length instabilities before and after the main instabilities. Small instabilities before the main one delay it because part of the fault has to be loaded to failure more than once. These short instabilities might be analogous to foreshocks and aftershocks, but a careful study would require shorter fault segments in the numerical model and substantially more computer time for solution.

Comparison of theoretical and observed uplift

Theoretical uplift for the reference case is now compared with observed uplift from the Osaka-Kushimoto level line and nearby tide gages (Fig. 3) reported by Thatcher (1984). Theoretical uplift is calculated with respect to uplift at a position near the Sakai tide gage, which Thatcher (1984) used as an invariant reference. Between 1945 and 1978 the Sakai tide gage moved less than 50 mm relative to sea level. In the present model, uplift at the Sakai position is essentially the same as uplift of the hanging wall block at infinite distance from the fault. Thus the footwall block far from the fault descends steadily with respect to sea level, as in the models of Savage (1983), Thatcher and Rundle (1984), and Thatcher (1984).
Theoretical and observed uplift curves are ordered on Fig. 8 from top to bottom according to increasing time. Circles are levelling data, and triangles are tide gage data. Smooth heavy lines are theoretical uplift changes for the time period indicated. As mentioned above, model parameters have been adjusted to produce the recurrence interval of about 92 years and agreement between theoretical and observed uplift for the coseismic period. The heavy curve in Fig. 8b is the theoretical change from fifteen years before instability to one-half year after instability. The curve is a little broader than the observation curve, but the overall fit is acceptable. Lighter curves marked 0~ and 0+ are for changes from 15 years before instability to instants just before and just after instability. The 0~ curve indicates that elevation change in the model between the years 1929 and 1946 is small compared to the change during instability. On the other hand the O"1" curve indicates that almost as much theoretical change occurs during the half year after instability as during.
The remaining panels in Fig. 8 are independent checks of the model. Agreement
is good overall and comparable to that of Thatcher and Rundle (1984) except for the postseismic intervals (c) and (d). In (d) the right (inland) part of the theoretical curve

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greatly underestimates the observed subsidence. It is not possible to adjust model parameters without degrading the agreement for other time periods. The much better fit for the post-earthquake period of (d) in the model of Thatcher and Rundle (1984, Fig. 14) (fault dip = 30°) is traceable to viscoelastic relaxation of the mantle in their model, a feature absent in the present model. In Fig. 8c theory and observation are in good agreement except for the leftmost observation point. This observation implies zero or small slip on the upper part of the fault during the first year after the earthquake, whereas the upper part of the model fault, Fig. 5b, continues to slip for a year after instability. Fig. 9 shows the theoretical fault slip changes responsible for the uplift changes in Fig. 8. Each curve shows the variation of fault slip with position at the time marked; the times are the same as in Fig. 8. Fault slip during a time interval is the difference between two curves. It is easy to see, for example, that the maximum coseismic subsidence is due to fault slip above £ = 80 km (1947 minus 1929). Similarly, the uplift change during the interseismic period, Figs. 8e and 8a, is due to fault slip increasing with depth below £ = 50 km (1929 minus 1899, 1979 minus 1967).
Theoretical curves in Figs. 8 and 9 are consistent with the two time-scales of observed postseismic deformation noted by Thatcher (1984). The first time-scale is about a year long and is explained in the instability model by slippage near the lower end of the unstably slipped fault, i.e., the 'intermediate-term postseismic' slip in Fig. 5b. The 'modified elastic halfspace' model of Thatcher and Rundle (1984) explains the same phenomenon by a temporally decaying slip just beyond the lower end of the seismic fault.
The second time-scale in postseismic geodetic data noted by Thatcher (1984) lasts several decades and is a diffusion-like migration of deformation landward. Figs. 8d, 8e, and 8a show the general pattern of migration and decreasing amplitude of observed uplift maxima. The positions of model curve maxima also migrate while amplitudes diminish, but the positions of maxima in (d) and (e) are offset to the right, and the model fails to produce the observed landward subsidence in (e) and (a). In the model of Thatcher and Rundle (1984) viscoelastic relaxation accounts for the diffusion, whereas in the instability model the diffusion is due to stress relaxation associated with spreading of slip on the fault. Thus in the data of Fig. 8 viscoelasticity seems to be required only to explain uplift during 1947-1967. It may be possible, however, to improve the agreement between theory and observation without viscoelasticity by allowing fault dip to increase with depth.

Intermediate- and short-term precursors

I now turn to pre-instability fault slippage and uplift and their possible use in forecasting earthquakes. First I will discuss curves of uplift versus time for five hypothetical benchmarks during the time interval from immediately after instability 1 (Ansei II) to immediately before instability 2 (Nankaido) on Fig. 5. Then I will focus on short-term precursors and compare a theoretical uplift curve with a tide gage record.
The uplift curves in Fig. 10 show two time-scales for post- and pre-instability stages,
as well as the slowly varying motion between instabilities. Benchmark positions are chosen to illustrate the range of uplift changes, and curves are labelled by their distance from the fault trace (cf. Fig. 3). Benchmarks at distances 10 to 40 km correspond to positions offshore. Intermediate-term precursory motion appears as increasing curvature starting

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several years before instability 2. As expected from the fault slip distribution in Fig. 5, the sense of curvature near the fault trace is opposite the sense far from the trace. The benchmark at x = 10 km has the largest departure from the earlier trend (the light line), about 50 mm over 5 years. No field data are available on the intermediate-term time-scale for comparison.
Short-term precursors, which appear as short vertical segments at the right ends of curves in Fig. 10, are shown magnified in Fig. 11. All anomalies start about 5 days before instability and are caused by slip on the interval 0 < £ < 30 km (Fig. 5b). The benchmark at x 10 km has the greatest departure from the prior trend, about 100 mm. Both intermediate- and short-term anomalies for x < 50 km would be large enough to detect in field data if measurements were made on land.
In the context of shear cracks, intermediate- and short-term precursory faulting episodes have slightly different mechanisms. Referring to Figs. 5 and 6, it is seen that intermediate-term faulting is approximately uniform over the brittle region. Short-term precursory faulting, in contrast, is due to nucleation of rapid slip and subsequent propagation away from the nucleation site. Sometimes such nucleation and spreading occurs more than once, as before instability 3 in Fig. 6b, and may be unilateral or bilateral. When nucleation and spreading occurs, it decreases the amount of subsequent unstable slip. Similar precursory faulting occurs in simulations analyzed by Tse and Rice (1986) for cases of most negative (^4 B) m-m, but not for the least negative (^4 #)min-
As noted in the discussion of Figs. 6 and 7, intermediate- and short-term pre-instability faulting varies with (^4 B) mi n and L. Intermediate-term anomalies for cases (a)-(c) are of comparable duration and size, as Fig. 6 implies, though there is some variation of shape. Short-term anomalies for cases (a)-(c) have greater variation. Anomalies for (a) are about twice as long and large as for (b), and they have more complex shapes. Short-term anomalies for case (c), as indicated by Fig. 6c, essentially vanish; durations are several hours and amplitudes are less than 10 mm. For antiplane faulting Stuart and Mavko (1979) and Tse and Rice (1986) showed a similar dependence of pre-instability faulting on fault law stiffness, which here is measured by the ratio (^4 B) m\n /L.
A tide gage record in Sato (1977) and Mogi (1985b) shows a possible short-term precursor starting two days before the Nankaido earthquake. The top of Fig. 12 shows the data plotted as the difference between tides observed at Tosashimizu (T in Fig. 3) and at Hosojima, with corrections for tidal currents. Hosojima is taken as a reference because it had negligible change during the Nankaido earthquake, while Tosashimizu had a 500 mm uplift. The precursory uplift starting December 19 is a distinct departure from the preceding trend, though there are two similar but less persuasive excursions November 24-26 and December 11-13. The tide gage data is not directly comparable to theoretical uplift because Tosashimizu is located more than 250 km from the geodetic data used to calibrate the model, is near the southwest edge of seismic rupture, and is adjacent to the large rectangular dislocation of Ando (1975) instead of the small one used here. One way to compare the computed uplift with the tide gage data is to assume that Tosashimizu measures deformation due to precursory slip appropriate to the large rectangular area and that the model can be scaled to the larger rectangle. In this case the position of Tosashimizu scales to x 40 km (Fig. 11).

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The solid curve at the bottom of Fig. 12 shows the theoretical uplift for the benchmark
at x = 40 km plotted with the same axes as for the tide gage data. The observed and theoretical uplifts have comparable amplitudes and time-scales, but the slopes have opposite signs because the benchmarks are on opposite sides of the dislocation pileup at the downdip end of the short-term precursory slip (Fig. 5b). If the tide gage motion is to be explained by the instability model, pre-instability slip would have to be northwest of Tosashimizu. Such deep slip occurs in Fig. 6a, for example, and can be made to happen in other cases by adjustments to (A B)(£). Thus Fig. 12 should be viewed only as a qualitative check of the existence of short-term precursory uplift predicted by the model. Mogi (1985b) discusses the tide gage data further and also an uplift anomaly of comparable amplitude .and time-scale before the Tonankai earthquake.
In a related instability model for thrust faulting before the M=6.4 1971 San Fernando earthquake in southern California, Stuart (1979) found that accelerating fault slip started several years before instability. Uplift data of Castle et al. (1976) were in fair agreement with the theory, though there were too few data for a careful check. Thatcher (1976) had previously shown that the observed preseismic uplift could be explained by updip migration of edge dislocations.

Conclusions

The main purpose of this study is to determine if an instability model could have application to earthquake forecast attempts for the Nankai Trough and similar subduction zones. There are two main conclusions. First, the model is consistent with most of the major features of seismic and aseismic elevation changes associated with the Nankaido earthquake. Second, the anomalous preseismic uplift predicted by the model is large enough to detect by existing survey methods. Thus, even though another earthquake similar to the Tonankai and Nankaido events is not expected soon at the same locations, a similar instability model may have application to other subduction zones in the world where earthquakes might occur in the near future. A two-dimensional model like the one presented here is probably inadequate for application to the Tokai seismic gap (Mogi, 1985a), but construction of a three-dimensional model seems warrranted.

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X

Fig. 1 Geometry of model. Ground surface is at y = 0, a is fault dip, and constant fault slip rate Vp occurs below f = d.

Ui
U n
v

y

Fig. 2 Fault segments used for numerical solution.

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Fig. 3 Map of southwestern Japan showing Nankai Trough. Rect-angles are vertical projections to ground surface of disloca-
tion models by ANDO (1975) for the Tonankai and Nankaido earthquakes. Dots with vectors are epicenters and slip di-rections from KANAMORI (1972). Triangles are tide gage stations, and dashed lines are levelling routes. Only level
line Osaka-Kushimoto and nearby tide gages are used; other level lines and tide gages are plotted to indicate additional data discussed by THATCHER (1984). Figure modified from ANDO (1975) and THATCHER (1984).

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(a)

vo
ro

15

100 150

£, km

10

(b)

+0.9yr
+4d INTERMEDIATE-TERM
POSTSEISMIC

INTERMEDIATE-TERM
PRECURSOR

BRITTLE VISCOUS

50

£, km

100 150

Fig. 5 Theoretical fault slip versus down-dip distance for reference case, (a) Curves every 20 time steps plus immediately before
and after instability, (b) Curves at boundaries of stages near
the Nankaido earthquake.

CA-B) mm = -0.6 bar
L= 50 mm

LEAST UNSTABLE

CASE OF CaHb)

(a) CA-B) min = -0.8 bar REFERENCE CASE
L - 40 mm

(b)

Fig. 6 Theoretical fault slip versus down-dip distance for four cases (a)-(d) of increasing intensity of instability. Panel (b) is the reference case in Fig. 5.

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-0.4

CO
. Q

c
E
S -0-8

- 1.2

- 1.6

10 20 30 40

L, mm

50 60 70 80

Fig. 7 Plot showing values of (A -B) min and L used for simula-tions. Dots marked (a)-(d) represent simulations in Fig. 6.
Circled dot is reference case.

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2.Or

100 80 60 40 20 50

YEARS BEFORE INSTABILITY 2 (NANKAIDO)

Fig. 10 Theoretical uplift versus time for instability 1 (Ansei II) to instability 2 (Nankaido) for six positions on the ground
surface.

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PHYSICAL BASIS FOR THE MAGNITUDE CUT-OFF

DEPENDENCE OF SEISMICITY QUIESCENCE

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Abstract

Seismicity for a few years preceding some moderate California earthquakes was re-examined for possible precursory patterns. The area examined for each moderate earthquake is similar in size to its aftershock zone. We found that in many cases the pattern depends on threshold or cut-off magnitude for data selection. When the threshold magnitude is about 2, enhanced seismicity or increased seismicity rate is usually observed before the earthquake; when the threshold magnitude is raised to 3, quiescence or decreased seismicity rate is observed before the earthquake, which may serve an intermediate term precursor for earthquake prediction. We noticed similar phenomena reported in many earlier studies. We propose to explain the threshold or cut-off magnitude dependence of seismicity patterns before moderate earthquakes as a consequence of the slip-weakening friction law. In our previous work simulating seismicity with a slip-weakening friction law, we showed that quiescence of seismicity before a large earthquake occurs when the critical weakening slip is large, but not when it is small. A recent study by Aki on the frequency-magnitude relation for small earthquakes suggests distinctly different parameters of the friction law between earthquakes with magnitude greater than 3 and those smaller than 3, suggesting that the critical weakening slip may increase sharply with increasing magnitude at around 3. Thus, our numerical simulation can explain why the seismicity quiescence shows up only for earthquakes with magnitude greater than about 3.

INTRODUCTION

A period of quiescence of seismicity lasting for a few years has been proposed as a precursor to many large earthquakes (Inouye, 1965; Mogi, 1968; Kanamori, 1981; Fedotov, 1982). It is sometimes also called temporal gap as compared with spatial gap and is an important part to make up a doughnut seismicity pattern (Mogi, 1969; Kanamori, 1981). Seismic quiescence as a precursor for earthquake prediction has received increasing attention because of (1) intermediate precursor time, (2) relatively high success rate for earthquake prediction, and (3) possible relations between quiescence and the nucleation process of a large earthquake.

So far most studies about seismic quiescence have been concerned with the identification of quiescence in the observed seismicity (Kanamori, 1981; Habermann, 1981; Wyss and Habermann, 1979; Mikumo and Miyatake, 1983; Mogi, 1969, 1977; Habermann and Wyss, 1984, among others). In the identification of quiescence the threshold or cut-off magnitude dependence of seismic quiescence has bee noticed (Evison, 1977; Tsai et_ a]_., 1979), but no attempt has been made to explain why a certain cut-off magnitude is chosen for demonstrating a quiescence pattern. In this paper we shall show that the cut-off magnitude dependence of seismicity pattern may be fundamental to the understanding of physical mechanism of seismic quiescence.

In the following we shall first give a few examples from Southern
California to show that the precursory seismic quiescence is cut-off magnitude dependent. Then we reconfirm our earlier numerical seismicity simulations with a displacement hardening-softening friction law (Cao and Aki, 1984/85),

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in which the seismic quiescence was demonstrated only for a certain level of slip weakening displacement. With this result of numerical simulations we can explain the observed cut-off magnitude dependence of seismic quiescence if we notice that the slip weakening displacement is dependent on event size (magnitude) and does not obey the self-similar scaling. After a brief review of the observational results about the breakdown of the self-similar scaling, we propose that this breakdown is the reason for the cut-off magnitude dependence of seismicity quiescence.

Cut-off magnitude dependence of seismicity quiescence

Let us first examine a few major earthquakes in Southern California for which the seismicity pattern was studied earlier. These earthquakes are 1952 Kern County (M=7.7), 1971 San Fernando (M=6.4), and 1979 Imperial Valley (M=6.9). In fact, different conclusions have been reached at least for the seismicity patterns before the first two events. Wesson and Ellsworth (1973) found that in years preceding the above first two events and some other moderate earthquakes in California the level of small earthquake activity in the epicentral areas were relatively high. On the other hand, Ishida and Kanamori (1978, 1980) found these two events were preceded by seismicity quiescence. Let us re-examine the seismicity patterns preceding these earthquakes to resolve these seemingly contradictory conclusions.

We used the seismic catalog prepared by Caltech (Hileman ^t aj_., 1973,
and monthly reports for recent years). According to Hileman j?t_ ^1_., the detection threshold for White Wolf Fault Area and Los Angeles Area, where the 1952 Kern County and 1971 San Fernando earthquakes respectively belong to, was magnitude 3.0 during the time period from 1932 to 1971. For the Imperial Valley region the detection magnitude threshold was 3.5 during the time period from 1932 to 1971 but had been improved to 2.4 for several years before the 1979 event (Habermann and Wyss, 1984).
There was no significant detection or reporting changes for several tens
of years before the 1952 Kern County and 1971 San Fernando earthquakes and for several years before the 1979 Imperial Valley earthquake (Hileman et al., 1973; Habermann, 1983; Habermann and Wyss, 1984). There is, however, a need for removing the so-called dependent events in a catalog in order to secure unbiased identification of quiescence. Dependent events (aftershocks, swarms, foreshocks) has been recognized and removed from the catalog by various methods (Gardner and Knopoff, 1974; Savage, 1972; Schlien and Toksoz, 1972; McNally, 1976; Habermann, 1981; Reasenberg, 1985). In the present paper we use the algorithm due to Reasenberg (1985).

The basic idea of Reasenberg's method is to recognize dependent events according to an interaction zone, which is defined by the time and space intervals between events. Generally speaking two events occurred closely in space and time are likely to be dependent. The interaction zone size to relate two events is a function of magnitude. In Reasenberg's model, for examples, earthquakes with magnitude 2.0 and 4.0 will have interaction zone sizes 0.7 and 4.6 km, respectively, for the next event. Considering the

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location error which is sometimes 5km or larger (Hil-eman ^ jal_., 1973), we double the zone size to 1.4 and 9.2 km respectively.

Figure 1 shows the cumulative number of all events plotted against time
for the area (32.7°-33.4° N and 115.7°-116.4° W) containing the 1968 Borrego Mountain earthquake. We see that due to the dependent event (mostly aftershocks of few major events) the curve is very irregular. Figure 2 shows the result in which the dependent events have been removed. The number of independent events is only one third of the total. We see that for 40 years the seismicity rate is nearly a constant. The comparison between Figures 1 and 2 illustrates how the algorithm for removing dependent event works.

Let us now consider the 1952 Kern County earthquake. This earthquake occurred along a 50-km section of the north-east striking White Wolf fault with northwest thrust accompanied by left lateral strike slip. According to Wesson and Ellsworth (1973) and Rienter (1955) the 20 year data preceeding the main shock showed that the activity within about 15 km of the epicenter was substantially higher than the activity in most of the surrounding region. They also found that the other end of the aftershock zone than the end containing the mainshock epicenter was quiet before the main shock and active after the mainschock. In Figures 3 and 4 we plotted the cumulative number of events vs. time relation using the catalog from the area surrounded by 34.7°-35.5° N and 118.2°-119.5° W. Figure 3 is for the cut-off magnitude 2.0; Figure 4 is for the cut-off magnitude 3.5. In both cases, the dependent events have been removed from the catalog. We see that in the low cut-off magnitude case the seismicity rate is almost a constant for the 20 years before the mainshock. The most significant feature of this figure in comparison with Figure 4 is that the seismicity was not decreased but somewhat increased before the 1952 main shock. This result is consistent with Wesson and Ellsworth's analysis (1973). With a higher cut-off magnitude, Figure 4 shows an 12 year long seismic quiescence before the 1952 main shock. This result is similar to Ishida and Kanamori's analysis (1980). Their identified quiet period was 15 years before the main shock. They also noticed the relatively high activity in the immediate vicinity of the main shock in the same time period before the main shock.
For the 1971 San Fernando earthquake we choose the rectangular area centered at the main shock surrounded by 34.0°-34.5° N and 118.0°-118.7° W, which is slightly larger than the aftershock area. Figures 5 and 6 show the cumulative number of events plotted against time using the catalog with dependent events removed. The result for the cut-off magnitude 3.0 is shown in Figure 6, where we see a 3 year quiescence before the earthquake. During the same time period, the seismicity with cut-off magnitude 2.0 kept the long-term rate with a slight enhancement immediately before the mainsthock. Again both the seismicity enhancement and quiescence are confirmed respectively by Wesson and Ellsworth (1973) and Ishida and Kanamori (1978). Wesson and Ellsworth found that between 1963 and 1970 San Fernando area had three broad areas of activity. Ishida and Kanamori identified a quiet period between 1965 to 1969 and an enhanced period from December 1970 to February 1971 for M>1.5. This example again shows the cut-off magnitude dependence of seismicity pattern. In Figure 6, however, we also find three quiet periods comparable to the one preceding 1971 San Fernando earthquake. These quiet

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periods are from 1932 to 1936, 1943 to 1949, and 1954 to 1960. They are not followed by any major earthquake. The total number of events (M>3.0) in Figure 6 is 28, which is too small to make a statistically significant conclusion. Therefore, the possibility that all above four quiet periods are due to statistical fluctuation cannot be excluded.

The precursory seismic quiescence before the 1979 Imperial Valley
earthquake has been studied by Johnson and Mutton (1982) and Habermann and Wyss (1984). With the dependent events removed, Habermann and Wyss found that there was a 15-week quiet period prior to the main shock if the cut-off magnitude is >2.5. They noticed that no quiescence could be defined if all events (after dependent events removed) were included. Habermann, however, admitted that a period of 15 weeks may be too short to have statistical significance.

The most recent example for the cut-off magnitude dependence of
seismicity pattern before moderate eartho.:akes is the Oceanside earthquake (July 13, 1986, M=5.3). Hutton et_ _al_. (1986, written communication) plotted the seismicity before this earthquake with cut-off magnitudes M=2.5 and 3.0. Their figures show that the seismic activity increased for M>2.5 and decreased for M>3.0 before the main shock.

In fact, the cut-off magnitude dependence of seismicity pattern before
large earthquakes, or more precisely the enhancement of seismicity with low cut-off magnitude and the quiescence of seismicity with high cut-off magnitude, was found by several other investigators earlier. For example, Evison (1977) found enhanced seismicity followed by a precursory quiescence for the 1968 Inangahua, New Zealand, earthquake (M=7.1) with a cut-off magnitude M>4.5. When he included all the small recorded events the quiescence was no longer recognizable. Tsai ot_ a]_. (1979) studied the seismicity of microearthquakes in the 2.5-3.5 magnitude range before several moderate earthquakes (M 5.7-6.8) in eastern Taiwan. They found the precursory increase in the monthly frequency of occurrence of these microearthquakes. When they raised the cut-off magnitude the precursory seismicity enhancement disappeared. The reasons for such a cut-off magnitude dependence was not discussed by these authors. Now, we shall try to look for a physical explanation for the cut-off magnitude dependence.

Seismicity quiescence and fault slip weakening

The physical explanation of seismicity queiscence proposed earlier may be divided into four categories. The first attributes to a heterogeneous distribution of stress (Mogi, 1977) or strength (Tsumura, 1979; Kanamori, 1981); the second to the precursory or aseismic slip (Habermann, 1981; Cao and Aki, 1984/85); the third to stress corrosion accompanied with precursory aseismic fault slip (Ohnaka, 1985); the fourth to dilatancy hardening (Scholz, 1986). In the numerical simulations of Cao and Aki (1984/85), they showed that with a displacement hardening-softening friction law characterized by the critical slip weakening displacement of the order of one centimeter the seismicity quiescence before larger earthquakes can be simulated. They also

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simulated seismicity patterns with the same friction law but critical slip weakening displacement one order of magnitude smaller and larger than one centimeter. In the first case, the simulated seismicity showed gradual enhancement until a main shock occurs. In the second case, the simulated seismicity showed mainly repeated major events with very few small events in between. The major conclusion they reached is that the seismic quiescence depends on the critical weakening slip. Figure 7 is reproduced from Cao and Aki's (1984/85), in which they compared the simulated seismicity patterns using two different friction laws, the simple friction law (Figure 7a) and the slip weakening friction law (Figure 7b), but the same fault strength distribution and initial stress distribution. With the average critical slip weaken displacement of 0.8 cm (Figure 7b), they simulated a ten-year-long quiescent period before a major earthquake. But with the simple friction law (Figure 7a), which is equivalent to a zero critical displacement, they only got enhanced seismicity before the mainshock. Thus, different friction laws generate different seismicity patterns.

In order to further confirm the validity of our conclusions about seismic quiescence, here we repeat our simulation for a variety of parameters. The displacement hardening-softening friction law used is the simplified form of Stuart (1979a, b,) and Stuart and Mavko (1979), namely,

T = S exp[-(u)-u) 0) 2 /a 2]

where T is the fractional stress, w the fault slip, and S, u) 0 » and a are material constants. Here the exponential term describes the displacement hardening and softening response of fault friction or fault strength.

The one-dimensional block spring model originally due to Burridge and Knopoff (1967) is used in our simulations. In this model an array of discrete blocks represents the material adjacent to the fault and fault friction is replaced by the fractional contact between blocks and a plane. 50 blocks are used in the simulations.

The peak fractional stress S was randomly chosen for 50 blocks from the uniform probability between two selected values. We assumed that a and u) 0 are proportional to S. Furthermore, to make S about 10% higher than static friction T(OJ=O), we put a=5u) 0. After these assumptions, the parameters we can choose freely are the range of strength S, the ratio between S and a (or o) 0), and the initial condition of stress distribution along the one dimensional model (initial positions for 50 blocks). For a fixed S, larger a means larger critical slip weakening displacement. For each set of above three parameters, we select a new seed for generating random numbers. In the following we shall present results for four simulations.

In Figures 8 to 11 the vertical axis gives the locations of blocks along
the fault and horizontal axis indicates the time of seismic events (+) or the initiation of aseismic slip (o). The vertically consecutive symbols (+) represent a single seismic event extended over these neighboring blocks nearly simultaneously. If all 50 blocks slip together, it is a mainshock. In this study we are primarily interested in the seismicity patterns before these

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mainshocks. We chose the range of S from 100 to 800 bars; w0 and a range from 0.03-1.3 cm and 0.15-6.5 cm. According to our previous simulations (Cao and Aki, 1984/85), these ranges of w0 and a should lead to a seismicity pattern in which the enhanced activity was followed by a few years quiescence before a mainshock.

Figures 8 and 9 are two simulations with the same homogeneous initial
stress along the fault and same random number generator seed for specifying strength distributions, we changed two blocks (number 45 and 49) with high strengths in Figure 8 to two blocks with moderate strengths in Figure 9. This change makes the fault segment from blocks 40-50 in Figure 9 more homogeneous and slightly lower in average strength. We see the simulated seismicity in Figure 9 is slightly different from Figure 8. However, the major feature of these two simulations is the same, that is, the enhanced seismicity starting around T=82 years is followed by a 3-5 years quiet period before the mainshock.

Figures 10 and 11 are two simulations with different random number seeds
for strength distribution and initial stress along the fault. The initial stress in these two simulations are more heterogeneous than in Figures 8 and
9. We find that the simulated seismicity patterns in Figures 10 and 11 are very similar. Both have gradually enhanced seismicity followed by a quiet period of 5 years long. The simulated seismicity patterns in the first two examples are also very similar to the second two examples, suggesting that seismic quiescence may be an inevitable consequence of a displacement hardening-softening friction law with a certain range of critical weakening si ip.

Magnitude dependence of siip weakening displacement

On the basis of a specific barrier model (Das and Aki, 1977; Aki, 1979), Papageorgiou and Aki (1983) estimated the critical slip weakening displacements for several California earthquakes from observed strong motion acceleration power spectra. They found the ratio between slip weakening displacement and the cohesive zone size is about 1/1000. For these California earthquakes with a magnitude range between 6.5 to 8.25, the estimated slip weakening displacement ranges from 0.4 m to 4.0 m. These results have been confirmed by totally different approaches, in which the recurrence behavior of a fault is numerically simulated (Stuart, 1985; Mavko, 1983; Cao and Aki, 1984/85). In these numerical simulations, the magnitude of critical slip weakening displacement similar to the estimate by Papageorgiou and Aki was required in order to produce observed recurrence behavior in space, time and magnitude.

The estimation of critical slip weakening displacement has also been made
for stick-slip experiments in the laboratory. Okubo and Dieterich (1983), for example, showed that the critical slip weakening displacements are 5 to

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25x10"** cm for a 2 m long fault length. Although we have no estimate of critical displacement for event size between large earthquakes and laboratory experiments, we may speculate that the critical displacement will decrease from the order of 10 cm to the order of 10~ 3 cm by some step-like change with the decrease in event size. On the other hand, as discussed by Dietrich (1979), there should exists a minimum earthquake for the friction law with a fixed critical displacement. So we may expect a breakdown of the self similar magnitude -frequency relation at least at certain magnitude range. This problem of minimum magnitude was addressed recentely by Aki (1987). He found a kink in the frequency-magnitude relation at M=3 for small earthquakes observed in a borehole in the Newport-Inglewood fault zone, and suggested that the minimum magnitude for the fault zone may be M=3.

The source dimension of an earthquake with M=3 is a few hundred meters,
and is of the same order of magnitude as the size of cohesive zone estimated for major California earthquakes by Papageorgious and Aki (1983) from observed strong motion acceleration spectra. It is natural to expect a distinct difference in the parameters of friction law between cracks with the size greater and smaller than the size of cohesive zone. We expect that the earthquakes with M<3 will have a distinctly smaller cohesive zone and consequently smaller critical weakening slip than those with M>3. Then, our numerical experiments would predict that the seismicity quiescence may show up only in the seismicity pattern of events with M>3, but not in that with M<3.

DISCUSSION

We have shown some evidence for the cut-off magnitude dependence of precursory seismicity quiescence and proposed the slip weakening friction law as a possible physical basis for such a cut-off magnitude dependence. Our observation that the seismicity quiescence occurs for M>3 and the seismicity enhancement occurs for M<3, however, contradict the commonly accepted precursory behavior of b-value change before major earthquakes. Our conclusion leads to a precursory increase of b-values, while before many reported precursory b-value changes show decrease. In order to clarify this point, we examined an extensive data collected by Seggern _et_ aj_. (1981) instead of finding examples with any bias.

Table 1 includes all the cases listed by Seggern et_ a]_. (1981) for precursory seismicity quiescence with mainshock magnitude equal to or larger than 5. The reference for each earthquake can be found from Seggern et al. (1981). We have added the cut-off magnitude as the last column. It is extremely interesting from our point of view that all the cut-off magnitudes, except for 1975 Hawaii and 1969 Garm earthquakes, are equal to or greater than 3.0, although this may be also due to the detectability of seismic networks. Another equally interesting fact is shown in Table 2 which includes all cases of precursory b-value changes given in Seggern ot_ a]_. (1981). We have listed in the last three columns the cut-off magnitudes or magnitude ranges for deriving b-values, the b-value changes and the characteristics of sequences (foreshock, aftershock) from which b-values are obtained. As expected, we see

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from this table that most of the examples show b-value decreases before mainshocks. The only example of increase is from 1967 Caracas, Venezuela earthquake.

From Table 2, however, we also see that all the cass of precursory
b-value decrease are from comparisons between foreshock and aftershock sequence with very low cut-off magnitudes («3) or magnitude range below 3. The b-value increase expected due to quiescence for M>3 and enhancement for M<3 would be for a much wider magnitude range, including events with M>3. Furthermore, the precursor times in Table 2 for b-value decreases are very short in compare with the precursor times in table 1 for seismicity quiescence. This suggests that they may represent different physical processes or different stages of an earthquake preparation process. The only example, which has relatively long precursor time and is also from a cut-off magnitude as high as 3, is the 1967 Caracas earthquake. The precursory b-value change before this event is opposite to other examples and consistent with our working hypothesis. The precursory b-value increase was also reported by Smith (written communication) for all shallow New Zealand earthquakes of magnitude 6 or greater since 1965.
The fact that b-value decreases or increases depending on the magnitude range below or above about 3 was also noticed by Jin (personal communication) when she was studying the seismicity before the 1976 Tangshan, China earthquake (M = 7.8). She found that the precursory low b-value for the 2 year period before the mainshock was estimated using events smaller than M=3, while the high b-value for the normal period was estimated using events of magnitude range higher than 3. If the kink in magnitude-frequency relation reported by Aki (1987) applies to China, the precursory low b-value may be explained by the different magnitude range used for estimating b-value, thus not conflicting our working hypothesis.

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Table 1. examples of precursory seismicity quiescence

Main shock

1976 Central Alentians
1971 Central Alentians
1968 Borrego Mt., Calif.
1971 San Fernando, Calif.
1973 Oxnard, Calif.
1973 Michoacan, Mexico
1968 Oaxaca, Mexico
1970 Chiapas, Mexico
1967 Costa Rica
1975 Hawaii
1969 Garm, USSR
1968 Tokachi-Oki, Japan
1973 Colima, Mexico
1973 North Kuril Islands

Precursor time

MO
YR
YR
YR
MO
15 MO
2
12

YR
MO
14 MO
38 MO
18 MO
76 MO
22 MO
5 YR

Mainshock
Magnitude

Mb
Ms
Ms
Ms
Ms
Ms
Ms
Ms
Ms
Ms
Mb
Ms
Ms
Ms

= 5
= 7.1
= 6.4
= 6.5
= 5.2
= 7.5
= 7.1
= 7.3
= 6.2
- 7.2
= 5.3
= 7.9
7.4
= 7.3

Cut-off
Magnitude

4.5
4.5
3.0
3.0
3.0
3.0
4.0
4.0
4.0
2.0
1.5
5.0
4.5
4.5

Table 2. examples of percursory b-value change

Main shock

1960 Chilean
1967 Fairbanks, Alaska
1970 Danville, Calif.
1967 Caracas, Venezuela
1963 Kariba Lake, Zambia
1970 Fairbanks, Alaska
1966 Kremasta Lake, Greece

Precursor time

33 HR
MO
DA
YR
YR
DA
MO

Main shock
Magnitude

M = 8.3
M > 5
M L = 4.5
Ms = 6.6
M = 5.8
M = 3.0
M L = 5.9

Cut-off
Magnitude
or Magnitude
Range

0.6 - 3.0
1.5 - 2.5
0.8
3.0
1.4

2.4 - 4.2

b-value
Change

D
D
D
I
D
D
D

Sequences

F/A
F/A
F/A
13 yrs. data
F/A
F/A
F/A

D: decrease
I: increase

F/A: precursory b-value change was obtained by comparing foreshock sequence with aftershock sequence.

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1952 KERN COUNTY

200

C M>2.0 U

I I I I i I I

100-

VO
*>
VO

')9SI
TIME

Fig 3.

The cummulative number of earthquakes vs. time relation for the 1952 Kern County earthquake area. All the dependent events have been removed. The cut-off magnitude is 2.0. No seis-mic quiescence can be seen before 1952 mainshock.

The Response of Creeping Parts of the San Andreas Fault to Earthquakes on Nearby Faults: Two Examples

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Abstract

Rates of shallow slip on creeping sections of the San Andreas fault have been perturbed on a number of occasions by earthquakes occurring on nearby faults. One example of such perturbations occurred during the 26 January 1986 magnitude 5.3 Tres Pinos earthquake located about 10 km southeast of Hollister, California. Seven creepmeters on the San Andreas fault showed creep steps either during or soon after the shock. Both left-lateral (LL) and right-lateral (RL) steps were observed. A rectangular dislocation in an elastic half-space was used to model the coseismic fault offset at the hypocenter. For a model based on the preliminary focal mechanism, the predicted changes in static shear stress on the plane of the San Andreas fault agreed in sense (LL or RL) with the observed slip directions at all seven meters; for a model based on a refined focal mechanism, six of the seven meters showed the correct sense of motion. Two possible explanations for such coseismic and postseismic steps are (1) that slip was triggered by the earthquake shaking or (2) that slip occurred in response to the changes in static stress fields accompanying the earthquake. In the Tres Pinos example, the observed steps may have been of both the triggered and responsive kinds. A second example is provided by the 2 May 1983 magnitude 6.7 Coalinga earthquake, which profoundly altered slip rates at five creepmeters on the San Andreas fault for a period of months to years. The XMMl meter 9 km northwest of Parkfield, California recorded LL creep for more than a year after the event. To simulate the temporal behavior of the XMMl meter and to view the stress perturbation provided by the Coalinga earthquake in the context of steady-state deformation on the San Andreas fault, a simple time-evolving dislocation model was constructed. The model was driven by a single long vertical dislocation below 15 km in depth, that was forced to slip at 35 mm/yr in a RL sense. A dislocation element placed in the seismogenic layer under XMMl was given a finite breaking strength of sufficient magnitude to produce a Parkfield-like earthquake every 22 years. When stress changes equivalent to a Coalinga earthquake were superposed on the model running in a steady state mode, the effect was to make a segment under XMMl, that could slip in a linear viscous fashion, creep LL and to delay the onset of the next Parkfield-like earthquake by a year or more. If static stress changes imposed by earthquakes off the San Andreas can indeed advance or delay earthquakes on the San Andreas by months or years, then such changes must be considered in intermediate-term prediction efforts.

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Introduction

A number of moderate earthquakes on faults adjacent to the 200-km long creep-ing part of the San Andreas fault in central California have distinctly perturbed the rates of surface slip on strands of the San Andreas fault. Mavko (1982), Mavko and others (1985), and Schulz and others (1987) discuss some recent observations of this sort. The best example is the magnitude 6.7 Coalinga earthquake of May 2, 1983, which produced coseismic steps and profoundly aifected the slip rates at a number of creepmeters for months to years after the event. Such perturbations may offer opportunities to examine the response of the San Andreas fault system to reasonably well calibrated, externally imposed stress changes. Ideally, such perturbations might tell us something about the state of stress existing on the fault plane, and provide some clues to the mechanical laws governing steadily creeping and stuck portions of the fault system. Earthquakes on the San Andreas fault itself have also aifected creep rates (e.g., Burford and others, 1973; Nason, 1973; King and others, 1977) and also offer opportunities to study fault properties, but with a more limited class of imposed stress patterns.
Several hypotheses have been offered to explain slip observed on one fault soon after an earthquake on a different fault (e.g., Alien and others, 1972; King and others, 1977). One possibility is that the observed slip is triggered by the dynamic stress field associated with the earthquake shaking, which releases stresses that already were acting on the fault but which, perhaps, had not yet reached high enough levels to begin a creep event. A second possibility is that the observed slip is the response of the fault, acting like a sensitive strain meter, to changes in the static stress field accompanying the earthquake.
Alien and others (1972) concluded that displacements on the Imperial, Super-stition Hills, and San Andreas faults after the 9 April 1968 Borrego Mountain earth-quake were triggered in part because the inferred static stress changes should have produced a LL displacement on the San Andreas fault, whereas RL offset was ob-served. Triggered slip has also been reported on the northern San Andreas fault after the 24 April 1984 Morgan Hill earthquake, for example, (Schulz, 1984) and on the southern San Andreas fault after the 8 July 1986 North Palm Springs earthquake (Williams and others, 1986; Fagerson and others, 1986).
King and others (1977) proposed that coseismic steps observed on creepmeters along the San Andreas fault near Hollister could be explained as responses to static stress changes greatly amplified by a compliant fault zone in which elastic strains would be concentrated due to the material weakness of the zone. As a result, creep-meters in the zone might act at times like extremely sensitive strain meters. Wesson and others (1986) report apparent responsive slip on the Banning strand of the San Andreas fault after the North Palm Springs earthquake.
Understanding earthquake-induced creep perturbations could be important to intermediate-term earthquake prediction efforts. The simple models described in

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this report suggest that stress changes accompanying some larger earthquakes could retard or advance earthquake activity on the San Andreas fault by months or years which falls within the span of intermediate term predictions.

In the following sections, we present two examples of earthquakes located within tens of kilometers of the San Andreas fault that perturbed slip rates at creepmeter sites on the San Andreas fault. The first example is provided by the Tres Pinos earthquake which occurred close to the northern reach of the creeping section near Hollister, California. The second is the Coalinga shock which occurred adjacent to the southern end of the creeping section near Parkfield (fig. 1). Studies of earlier creep events near the sites of these earthquakes have been reported most recently by Evans and others (1981) and by Goulty and Oilman (1978).

Approach

Our studies are based on simple models using dislocations on rectangular sur-faces in an elastic half-space (e.g. Steketee, 1958; Chinnery 1966a, 1966b; Erickson, 1986). Although the earth is certainly not a homogeneous elastic half-space (e.g. Jennings, 1977), we anticipate that important lessons remain to be learned from naive, first-order models. Our ignorance of many important variables and processes within the earth makes simplicity a necessity if not a virtue. Such models should be interpreted with great caution only results of considerable robustness should be taken seriously. That is to say, believable results must not be sensitive to geometry or choice of model parameters and must be considered primarily as order-of-magnitude guides that can help foster a better intuition as to how faults are interacting in three dimensions.

If a simple model can make robust, order-of-magnitude predictions with an absolute minimum of assumptions, then perhaps the model includes the essential variables controlling the phenomenon being investigated. It is hoped that the exam-ples given here will illustrate this modeling philosophy, and will convince the reader of its utility in some applications.

Tres Pinos Earthquake

The magnitude 5.3 Tres Pinos earthquake occurred on January 26, 1986 about
10 km southeast of Hollister, California (fig. 1). The hypocentral coordinates are 121° 16.97' W, 36°48.21' N, and a depth of 7.5 km. The focal mechanism was initially determined to be right-lateral (RL) strike-slip on a steeply dipping plane striking slightly west of north (J. Eaton, oral commun., 1986). Subsequent refinements pro-duced a fault plane striking N 7°W, dipping 83° E, with an auxiliary plane at N 80°E dipping 75°N (P. Reasenberg, oral commun., 1986). Seven creepmeters on the San Andreas fault (fig. 2) showed creep steps either during or soon after the shock (table 1). Steps as small as three telemetered unit-counts (about 0.02 mm) can be reliably identified on these creepmeters. Because the telemetered data consists of one observation every 10 minutes, the time elapsed between the earthquake and the

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occurrence of a creepmeter step is known to only 10 minute accuracy. All of the steps listed in table 1 will be referred to in the discussion that follows as coseismic.

The static stress changes produced by the earthquake were modeled by RL slip of 3 m on a vertical 1-km-square dislocation striking N5°W positioned at the hypocenter. The hypocenter of the earthquake is about 11 km from the San Andreas fault plane, so to first order the dislocation can be considered a point source, which is why a 1-km size was chosen. This orientation was based on the preliminary focal mechanism. In this model, as in all subsequent models, Poisson's ratio was 0.25 and the rigidity was 3 x 105 bar. The predicted change in shear stress on the plane of the San Andreas fault, which is assumed to strike N 50° W in this vicinity, agrees in sense with the observed slip directions at all seven sites (fig. 3a). A model based on the more refined focal mechanism (fig. 4) rotates the lobes in the shear stress pattern slightly so that the observed sense of motion is no longer consistent with the predicted sense of shear stress at XFLl (fig. 3b).

In the case of the Tres Pinos earthquake it appears that both triggered and re-sponsive creep may have been recorded. The observed LL steps, if not an instrument artifact, could provide a critical piece of evidence favoring responsive creep. There are, however, two difficulties. The LL steps might reflect an instrumental problem, because the orientation of almost all creepmeters on the San Andreas fault relative to the fault trace is such that apparent LL steps will result from any stretching of the creepmeter wires during seismic shaking. (See Schulz and others (1982) for ref-erences on the designs and locations of the creepmeters.) This bias could be solved by constructing some of the meters to cross the fault in the opposite direction. The second problem is that the meter XSJ2 has frequently shown LL creep during the rainy season, probably because of near-surface instability at the site. It can be seen from figure 5 that several LL creep events occurred at this meter in the months be-fore and after January 26, 1986. Other LL events appear in January 1985. Thus it is possible that the LL coseismic step at this meter was entirely triggered: the meter was in a period of LL creep that was simply accelerated by the shaking. It does not seem possible to decide from the information available whether or not the coseismic slip at this site could have been a mixture of triggered and responsive types.
The long term record for XFLl, which also had a LL coseismic step, does not indicate any obvious prior LL tendencies (fig. 5). In the months after the earthquake, the meter showed a markedly slower rate of creep. If the slow-down in creep following the earthquake was caused by a reduction in static stress at the time of the earthquake, then the near-surface part of the San Andreas fault in this vicinity may be capable of responding to imposed changes in static stress fields on the order of tenths of bars. The existing ambient RL stress near the surface (at least at XFLl) would also have to be less than tenths of bars if imposed LL stress of this size could reverse them. Such stress amplitudes are consistent with theoretical predictions of stress changes accompanying shallow creep events made by Nason and Weertman (1973) and by R.L. Wesson (written commun., 1986).

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Evidence for triggered slip caused by the Tres Pinos event can be seen in the longer term behavior of the meters CWNl and CWC3. Creep events at these sites occurred with some regularity during the preceding years, at least compared with the irregularities apparent on the records of other meters (see Burford and others, this volume). In figure 5, it appears that a creep event that might normally have occurred at these two meters within several months was triggered early by the the Tres Pinos shock. Subsequently, there was a period with little or no creep that lasted longer than the two. episodes with little creep that preceded the earthquake triggered event (fig. 5). Although the coseismic creep event may have occurred prematurely because of triggering, its amplitude appears to have been as large or larger than the preceding events, perhaps because of added responsive RL slip.

It is possible to estimate the depth to which the fault might be adjusting coseis-mically. We assume a limiting case of the compliant fault zone model namely that the fault is a frictionless cut or crack in the surface of the elastic half-space to some depth and that the two sides will move independently in response to the changes in stress so that shear stresses across the crack will remain zero. A three-dimensional

approximation to a frictionless cut in the surface of the half-space was formed by alining 1-km long by 0.5-km deep dislocations along the trace of the San Andreas fault (fig. 6) and by letting them slip in response to the stresses generated by offset on the Tres Pinos dislocation so that the shear stress across each dislocation returned

to zero. As can be seen in figure 7, the amplitudes of the larger RL steps exceed by a factor of 5 to 10 the offsets expected for a cut of this depth. This observation is consistent with the behavior of creepmeters during successive aftershocks of the Coalinga earthquake (Mavko and others, 1985); certain meters that eventually re-sponded with LL coseismic steps tended to respond initially with large RL coseismic steps to the main shock and to initial aftershocks as if there was a backlog of RL slip that was being shaken out of the system.

In order to investigate whether changes in stress had affected the steady state behavior at greater depths, one of us (Dietz) examined the seismicity on the San An-dreas fault plane for one year proceeding and three months following the Tres Pinos main shock. The seismicity above magnitude 1.0 during this 15 month period was declustered using Reasenberg's algorithm (1985), and seismicity rates were analyzed following the method developed by Matthews and Reasenberg (this volume). No changes in the patterns or rates of seismicity could be perceived, mostly because the seismicity in this area was too sparse to establish any patterns at all. For example, the segment of the San Andreas fault which experienced the greatest stress change (> 0.1 bar, fig. 8) coincides almost exactly with the Cienega Winery Seismic Gap (volume 372 of Wyss and Burford, 1985) where the background seismicity rate is known to be low. It was also in this volume that we anticipated the greatest possi-bility of noticing any changes in the seismicity on the San Andreas fault following the Tres Pinos event. However, any conclusion drawn from the number of earth-quakes occurring here (8 events above Ml.5 and 16 above Ml.O within the 15 month period) cannot be significant. Figure 8 shows the predicted changes in shear and

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normal stresses and the increment to the Coulomb failure criterion. This increment is calculated by applying the change in horizontal shear stress to an assumed ambient RL shear stress field. Because the Coulomb failure criterion is not linear by virtue of an absolute value sign around the shear stress term, this increment is accurate only where the sense of the total shear stress does not change sign. This condition may be violated near the earth's surface at sites where the LL steps occurred.

We are presently investigating in collaboration with Chi-Yu King the patterns
of creep on the San Andreas in response to earlier earthquakes in the vicinity of Hollister. Initial results from three additional earthquakes have not shown the same same degree of consistency between observed slips and predicted shear as found for the Tres Pinos earthquake, although left-lateral coseismic and post-seismic offsets have been recorded. This suggests that the agreement between observed and pre-dicted slips might be fortuitous in the Tres Pinos case, and that triggered slip may in many cases mask weaker responsive slip.

Coalinga Earthquake

The magnitude 6.7 Coalinga earthquake of May 2, 1983 had a profound effect on creep rates recorded on the San Andreas fault (fig. 9 and figure in Burford and others, this volume). The creepmeter XMMl at Middle Mountain 10 km northwest of the town of Parkfield recorded LL slip for over one year after this shock with many LL coseismic steps accompanying individual aftershocks. Strain meter CLSl slightly to the south of XMMl showed persistent extension during the same interval, including coseismic steps that in many cases correlated with events seen on XMMl (Alan Jones, oral commun. 1986). Changes in microseismicity along the San Andreas fault are reported by Poley and others (1987), that also correlate spatially and temporally with anomalous creep observed after the Coalinga event.

Mavko and others (1985) summarized the effects of the Coalinga event and its aftershocks on the creep records, and calculated the shear and normal stresses im-posed on the San Andreas fault at the earth's surface by the Coalinga earthquake offset. Figure 10 shows the changes in shear and normal stresses on the plane of the San Andreas calculated for a dislocation surface striking N 53° W, dipping 23°SW, and centered at a depth of 6 km under the Coalinga epicenter. Horizontal length of the dislocation was 16 km, down-dip length was 8 km, and the reverse slip was 1.8 m. As Mavko and others (1985) demonstrated, the far-field elastic effects are indepen-dent of which of the two focal mechanism planes is chosen as the fault plane. Figure 10 also shows the incremental change in the Coulomb failure criterion calculated using only the change in the horizontal component of shear stress, which is assumed to be superposed on a large ambient RL shear stress. The same caution applies as before because the Coulomb criterion is not linear, this increment is meaningful only where shear stresses do not change sign, which does not appear to be true at the earth's surface at the XMMl site, for example. If, for the sake of argument, the shear stress on the fault near the earth's surface was everywhere zero before the

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Coalinga mainshock, then the absolute value of the shear-stress term would need to be determined before combining it with the incremental normal stress in Figure 7 of Mavko and others (1985, p. 486), which would shift the plotted values in their Fig-ure 7 toward the south. It is of considerable interest that apparent imposed stresses on the order of 1 to 2 bars were capable of producing such long-term changes in the shallow slip behavior on the San Andreas fault. This point will be addressed in more detail below.

In an effort to simulate the temporal behavior of the XMMl meter and to view the stress perturbation provided by the Coalinga earthquake in the context of the steady state stress field driving the San Andreas fault, two simple time-dependent dislocation models were constructed (fig. 11). The models were driven by displacements on a single long vertical rectangular dislocation surface extending from 15 km to 115 km in depth, that was forced to slip at 35 inm/yr in a RL sense. (The exact height of this dislocation is not very critical.) A set of 10-km square vertical dislocations extending from 5 to 15 km simulated the seismogenic layer, and a set of 5-km square dislocations just below the earth's surface, that could slip in a linear viscous fashion in response to shear stresses, simulated a near-surface creeping layer. A single rectangular dislocation surface in the seismogenic layer under XMMl was given a finite breaking strength of sufficient magnitude to produce a Parkfield-like earthquake every 22 years. Ten dislocations in the seismogenic layer to the northwest were allowed to slip viscously, to approximate the behavior of the creeping section of the fault between Parkfield and San Juan Bautista. In one model, one dislocation in this layer immediately to the southeast of XMMl was allowed to creep, whereas in the second model two additional dislocations with finite breaking strength were placed to the southeast. The fault plane farther to the southeast was held permanently fixed to simulate the locked part of the San Andreas fault southeast of Cholame Valley that last broke in the great Fort Tejon earthquake of 1857 (Sieh, 1978).

Viscously creeping segments respond only to shear stresses in these models, mostly because the effects of normal stress on creeping parts of the fault are not known and thus not taken into account. Locked segments with finite breaking strengths (cohesions) and frictions of 0.5 were governed by application of the Coulomb failure criterion. Failed segments were allowed to slip so that the shear stresses acting on them drop to zero. The interaction of the dislocations was calculated using the displacement-discontinuity boundary element approach described by Crouch (1976) and applied by Mavko (1982) in a two-dimensional formulation to study the inter-actions of faults in the Hollister region.
When stress changes equivalent to a Coalinga earthquake were superposed on the models running in a steady state mode (35 mm/yr slip at depth, the Parkfield segment breaking every 22 years), the effect was to make the creeping segment un-der XMMl slip in a LL sense for about year if the coefficient of linear viscosity were properly adjusted. This is not a delicate adjustment; however, if the viscously creep-

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ing dislocations are made frictionless so that they respond immediately to imposed stresses, then the segment under XMMl will not move LL, because the LL tenden-cies are swamped by the imposed RL impulse applied by the Coalinga offset farther to the north, which is transferred down the fault by nearest-neighbor interactions. Figure 12 shows the slip for the 5-km-square dislocations just below the surface, and figure 13 shows the shear stresses acting on the three 10-km-square locked segments south of Middle Mountain in the second model. In these plots, a net slip of the Coalinga dislocation equal to twice the inferred seismic slip of 1.8 m was used to make the changes more visible. The net effect of the Coalinga-like impulse on this simple model is to delay the onset of the next Parkfield-like earthquake by a year or more (provided that slip, including aseismic slip, in the Coalinga region is not a part of every Parkfield cycle). The magnitude 5.5 New Idria earthquake on October 25, 1982 and the magnitude 5.5 Kettleman Hills earthquake of August 4, 1985 also produced responses on the creepmeter records (fig. 9). The effects on the plane of the San Andreas fault predicted for these earthquakes by the dislocation models are very much smaller than the effects predicted for the Coalinga earthquake (fig. 14) if magnitudes are converted to slips on the fault planes using a moment-magnitude relation Iog10 M0 = 1.5M+ 16.0. (Ekstrom (1986) reports a moment of 1.7 x 1025 dyne-cm for the Kettleman Hills event from very broad band teleseismic analysis, which imply a considerably larger slip than the moment of 1.8 x 1024 dyne-cm used in these calculations.) The New Idria offset tends to retard the next Parkfield-like shock, but the Kettleman Hills earthquake, because it is farther to the south, pro-duces a RL shear near XMMl and tends to advance the Parkfield-like break in time. Given the likely presence of aseismic creep in the vicinity of Coalinga and Kettleman Hills both before and after the earthquakes, it is difficult to assess the net effect of all the slip in this area on the timing of the next Parkfield shock. It is interesting to speculate that the apparent accelerated RL slip that was observed on the XMMl and WKRl (and perhaps XPKl) creepmeters in the months before the Coalinga main shock, although it might have been caused by rainfall, could have been caused by aseismic slip in the vicinity of Kettleman Hills on a slip plane similar in orienta-tion to the earthquake fault plane. Such slip might have loaded the Coalinga fault and ultimately triggered the Coalinga earthquake. Erickson (1986) has examined some possible triggering effects of the Coalinga earthquake on the Kettleman Hills earthquake.

It is still surprising that the small imposed stresses in figure 10 could have such
a marked long-term effect on the shallow creep, and several additional possibilities need to be considered in this connection:

(1) The time duration of the left-lateral creep at XMMl, although simulated to first-order by the simple linear viscous models described here, might be driven by post-seismic slip on the Coalinga fault or on neighboring faults in that vicinity. In this case, the observed perturbation at XMMl would not be the response to a step impulse, but to a more gradual impulse distributed over time. This possibility is strengthened by the succession of left-lateral coseismic steps observed at XMMl in

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response to aftershocks of the Coalinga mainshock that were greater in magnitude than about 3.5.

(2) Distances measured repeatedly with a two-color laser to sites on both sides
of the San Andreas fault from a central station at Car Hill about 1.5 km south of Parkfield suggest that multiple strands of the fault system are active over time, and that alternation of slip from one strand to another may occur in a quasi-periodic fashion over a period of months. If this is so, then slip recorded at a single creepmeter may not tell the whole story. From July 1985 to June 1986, during a period of slowed RL and/or LL slip at XTA1 and XDR2 creepmeters, an alinement array (RCW5) on Ranchita Canyon Road was observed to be slipping RL at an average rate of about 15 mm/yr. This alinement array spans the Southwest Fracture, which runs subparallel to the San Andreas fault about 1.5 km to the southwest and was observed to have RL offsets following the 1966 Parkfield earthquake (Brown and Vedder, 1967). In August 1986, after the adjacent main-fault creepmeters resumed more normal RL motion, the alinement array was remeasured and was found to have taken a LL step after its nearly year-long RL excursion. Thus LL motions on one strand may accompany RL slip on adjacent subparallel strands.
(3) Geologic cross-sections in the vicinity of Coalinga and the west side of the Great Valley, constructed from multiple seismic reflection profiles and other geologic and geophysical data, show subhorizontal and shallowly dipping reflectors just one crustal thickness away from the San Andreas fault (C. Wentworth, oral commun., 1986). The existence of low-angle structures offers the possibility of extending fault offsets resulting from Coalinga-like earthquakes to positions considerably closer to the San Andreas fault, if not right up to the fault plane. Although all of these possibilities remain difficult to evaluate at this time because of a scarcity of data, their possible role in amplifying or changing the pattern of shallow creep recorded on the San Andreas fault must be kept in mind.

Conclusions

It is not yet possible to explain unambiguously the observed behavior of creep-meters on the San Andreas fault during and after earthquakes on other faults. How-ever, in several cases the behaviors can be simulated at least qualitatively by very simple models using dislocations in an elastic half-space. This suggests that creep observations following such earthquakes may help us to understand the rheology of the San Andreas fault zone and the distribution of stress in and around the fault

zone.

Acknowledgments

Our modeling efforts are based on equations derived by Glenn Converse and originally programed as DIS3D by Scott Dunbar. The program has been considerably revised and thoroughly documented by Laurie Erickson of Stanford University. These equations allow the calculation of elastic fields not just at the surface, but also at

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depth an essential ingredient if one wishes to look at the interaction of dislocations. We thank Jerry Eaton and Paul Reasenberg for providing information about the Tres Pinos focal mechanism. Much of the work reported here was inspired by and follows from the efforts of Gary Mavko.

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Table 1. Creepmeter Retpontet to the Tret Pinot Earthquake

Creepmeter

XSJ2
XHR2
CWN1
CWC3
XFL1
XMR1
BIT1

Step Size l
(mm)

-0.05
-f2.68
-fl.67
+1.54
-0.05
+0.35
+0.33

Time after Earthquake
(minutes)

<10
20-40
<10
<10
30-50
0-20
90-110

Distance from Epicenter
(km)

22
14
12
12
16
25
52

Plus is right lateral, minus is left lateral.

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122' 121 120'

37'

36'

35'

San Juan Bautista

Hollister

^1/26/86 TRES PINOS
:. XHR2
CWN1.CWC3

10/25/82
NEW IDRIA

Figure 1. Location diagram showing selected creepmeter sites on the San Andreas fault in central California and epicenters of nearby earthquakes that are referred to in the text.

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SAN ANDREAS FAULT
*

CWN1CRU.CWC3CRU
- 0.00

25 km

0 km

SAN ANDREAS FAULT

XHR2CRL)
f
CWN1CRL),CWC3(RL)4

-25 km

25 km

0 km

- 25 km

Figure 3. Map view at the earth's surface of predicted static horizontal shear stress changes accompanying the Tres Pinos earthquake. Horizontal shear stress is resolved onto planes parallel to the San Andreas fault, which is approximately parallel to the left edge of the figure, (a) Model based on preliminary focal mechanism, (b) Model based on refined focal mechanism. Epicenter is located at the center of the figures. Tics along edges of the plots are at 2 km interval; contour interval is 0.025 bar with hachures on contours pointing toward low values. Negative shear is RL, positive is LL. Small triangles show creepmeter locations; thin line marks trace of San Andreas fault.

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A DILATATION
O COMPRESSION

Figure 4. Refined focal mechanism for the Tres Pinos earthquake (prepared by P. Reasenberg).

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XSJ2

XHR2

CWN1

CWC3

XFL1

XMR1

BIT1

J3 mm

~~~

Hollister Rainfall

CM

J

_J_*>JVLJ,
J 0.01 inches

1985 N* 1986

Figure 5. Slip records beginning on January 1, 1985 for creep meters affected by the January 26, 1986 Tres Pinos earthquake.

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San Andreas Fault

Tres Pinos hypocenter

Figure 6. Dislocation geometry used to simulate a frictionless crack forced to slip by the changes in static stress field associated with the 26 January 1986 Tres Pinos earthquake. The individual rectangular dislocations along the San Andreas fault are 1 km long and 0.5 km tall.

975

CREEPMETER STEPS - 26 JAN 1986

LU

CO

D_
LU
H
CO

-2

-3

LEFT
LATERAL

RIGHT
LATERAL

I l

XMR1 BIT1

- - CWC3

X CWN1

- L XHR2

NW

20 10 0 -10 -20

DISTANCE ALONG SAN ANDREAS FAULT CKM)

-50
SE

Figure 7. Observed creepmeter steps shown as vertical lines compared with predicted shallow slip on the San Andreas fault for a frictionless 0.5-km deep crack simulated by rectangular dislocation surfaces.

976

HORIZONTAL SHEAR STRESS (+ IS LEFT LATERAL)

Q_
LU
Q

NORMAL STRESS C+ IS EXTENSIONAL)

INCREMENT TO COULOMB FAILURE CRITERION

C+ IS CLOSER TO FAILURE)

Figure 8. Predicted changes in (a) static horizontal shear stress, (b) normal stress, and (c) Coulomb failure criterion on a vertical plane striking N50°W in the approximate location of the San Andreas fault. The changes result from slip on a rectangular dislocation surface simulating slip accompany-ing the Tres Pinos earthquake, which occurred about 11 km to the northeast of the vertical section displayed here. Small rectangle is the projection of the dislocation surface onto the section. See caution in text about non-linearity of Coulomb criterion. Contour interval is 0.025 bar.

977

NW
0 km

15

PARKFIELD

XMM1 SE
55 km

HORIZONTAL SHEAR STRESS (+ IS LEFT LATERAL)

NORMAL STRESS (+ IS EXTENSIONAL)

15
INCREMENT TO COULOMB FAILURE CRITERION

B- IS CLOSER TO FAILURE)

Figure 10. Predicted changes in (a) static horizontal shear stress, (b) normal stress, and (c) Coulomb failure criterion on the plane of the San Andreas fault resulting from the Coalinga earth-quake. See caution in text about non-linearity of Coulomb criterion.

979

Ca) XMM1 MODEL

t

Cb) XMM1 MODEL

\

^

10

L

11

1

12 13

1-2

14 15

L 3

|

5 km

L
» 10 km

10 km

Figure 11. Geometry of dislocations for model 1 (a) and model 2 (b) seen projected onto a vertical section parallel to the San Andreas fault. Only the south end of the models near the XMMl creepmeter are shown. Segments marked with an L are locked with a finite breaking strength, except for the long area on the far right which never breaks; other patches creep in a linear viscous fashion. Parallelogram is the dipping dislocation representing the Coalinga earthquake as seen projected onto the plane of the San Andreas fault.

980

vo
oo

HORIZONTAL SLIP
COALINGA-LIKE

1940 1950 1960 1990 2000

Figure 12. Predicted horizontal slip for shallow 5-km-square dislocations in the vicinity of the
XMMl creepmeter in model 2, showing the effects of Parkfield-like earthquakes normally occur-
ring at 22 year intervals and of a simulated Coalinga-like earthquake in 1983. Numbers refer to
dislocations in figure lib.

SQ.00-^

Figure 14. Map view at earth's surface showing horizontal shear stress changes resolved onto planes parallel to the San Andreas fault for earthquakes at (a) New Idria, (b) Coalinga, and (c) Kettleman Hills. Contour interval is 0.025 bar for (a) and (c), 0.5 bar for (b). Slip used for (c) is probably an underestimate (see Ekstrom, 1986).

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Summary of session on Physical Models

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Most of the modeling work involves some form of precursory critical slip Au associated with stress-displacement softening, which can be either time, displacement, or velocity dependent. As shown in the figure, there is generally a direct relation between the critical slip in the stress displacement domain and the occurrence and duration of precursory deformation versus time.
It appears that the details of the stress-slip function very much control the type, style and presence of precursory deformation before the onset of instability. For example, the absence of an inflection point in the a u behavior leads to the absence of any short or moderate term precursory deformation.
A major challenge for these models is the discrepancy between critical slip length as observed in the lab (Au = 10~3 10~2 cm) and the slip required insitu (Au = 10 103 cm). Part of the problem may reside in the fact that the actual fault is geometrically very complex in comparison with the almost idealized lab analogue (e.g., Simpson), with many strands and slip which is transferred with time from one strand to another, such as documented in the Parkfield-Cholame area, and other pull aparts elsewhere. Furthermore, the precursory slip models do not as yet include heterogeneities in general although heterogeneities are not only ubiquitous, but probably central to dynamic faulting. Such heterogeneities include not only many strands and soft fault rocks but also pore pressure, frictional heating, and the like.
Finally, it is my opinion that these models have not as yet been able to account for rupture length - as opposed to slip magnitude. In a way it is rupture length, or the stopping of rupture which is the feature that needs to be predicted - since rupture initiations occur all the time everywhere in the form of small and very small earthquakes, as indicated by the frequency - magnitude relation.
The conclusion I draw from these considerations is that we are in the beginning stage
of understanding and simulating the earthquake process - so that a lot of debate and excitement lies ahead.

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PAPERS UNAVAILABLE FOR PUBLICATION

The following papers -were part of the conference program but were not available for publication.

Time-space scaling and repeatability of seismic precursors
for intermediate-term predictions of large (Ma > 7)
California earthquakes
K. McNally

Magnitude residuals before the 1979 Coyote Lake, California, earthquake
C. Michaelson and W. Bakun

Summary of session on Electrical, Magnetic, and Seismic Precursors
C. H. Scholz

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Conference II

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Conference XII

Conference XIII

Workshop XIV

Workshop XV

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Workshop XVII

Workshop XVIII

CONFERENCES TO DATE

Abnormal Animal Behavior Prior to Earthquakes, I Not Open-Filed
Experimental Studies of Rock Friction with Application to Earthquake Prediction
Not Open-Filed
Fault Mechanics and Its Relation to Earthquake Prediction
Open-File No. 78-380
Use of Volunteers in the Earthquake Hazards Reduction Program
Open-File No. 78-336
Communicating Earthquake Hazard Reduction Information
Open-File No. 78-933
Methodology for Identifying Seismic Gaps and Soon-to-Break Gaps
Open-File No. 78-943
Stress and Strain Measurements Related to Earthquake Prediction
Open-File No. 79-370
Analysis of Actual Fault Zones in Bedrock Open-File No. 79-1239
Magnitude of Deviatoric Stresses in the Earth's Crust and Upper Mantle
Open-File No. 80-625
Earthquake Hazards Along the Wasatch and Sierra-Nevada Frontal Fault Zones
Open-File No. 80-801
Abnormal Animal Behavior Prior to Earthquakes, Open-File No. 80-453
Earthquake Prediction Information
Open-File No. 80-843
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the Puget Sound Region,

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The Dynamic Characteristics of Faulting
Inferred from Recording of Strong Ground Motion Open-File No. 82-591
Hydraulic Fracturing Stress Measurements Open-File No. 82-1075
Continuing Actions to Reduce Losses from Earthquakes in the Mississippi Valley Area Open-File No. 83-157

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Active Tectonic and Magraatic Processes Beneath Long Valley Caldera, Eastern California Open-File No. 84-939
The 1886 Charleston, South Carolina, Earthquake and its Implications for Today
Open-File No. 83-843
A Workshop on "Continuing Actions To Reduce Potential Losses from Future Earthquakes in the Northeastern United States"
Open-File No. 83-844
A Workshop on "Site-Specific Effects of Soil and Rock on Ground Motion and the Implications for Earthquake-Resistant Design"
Open-File No. 83-845
A Workshop on "Continuing Actions to Reduce Potential Losses from Future Earthquakes in Arkansas and Nearby States"
Open-File No. 83-846
A Workshop on "Geologic Hazards in Puerto Rico" Open-File No. 84-761
A Workshop on "Earthquakes Hazards in the Virgin Island Region"
Open-File No. 84-762
A Workshop on "Evaluating the Regional and Urban Earthquakes Hazards in Utah"
Open-File No. 84-763
Mechanics of the May 2, 1983 Coalinga Earthquake Open-File No. 85-44
On the Borah Peak, Idaho, Earthquake
Open-File No. 85-290
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Evaluation of Regional Urban Earthquake Hazards in risk in Alaska
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Future Direction in Evaluating Earth-quakes Hazard of Southern California Open-File No. 86-401
Earthquake Hazards in the Puget Sound Washington Area
Open-File No. 86-253
Probabilistic Earthquake Hazards Accessment Open-File No. 86-185
Earth Science Considerations for Earthquake Hazards Reducation in the Central United States Open-File No 86-425
Assessment of Geologic Hazards and Risk Puerto Rico Open-File No 87-008

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Physical and Observational Basis for Intermediate-term Earthquake Prediction
Open-File No 87-591
Earthquake Hazard Along Wasatch Front, Utah Open-File No 87-154
Directions in Paleoseismology
Open-File No.87-XXX
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