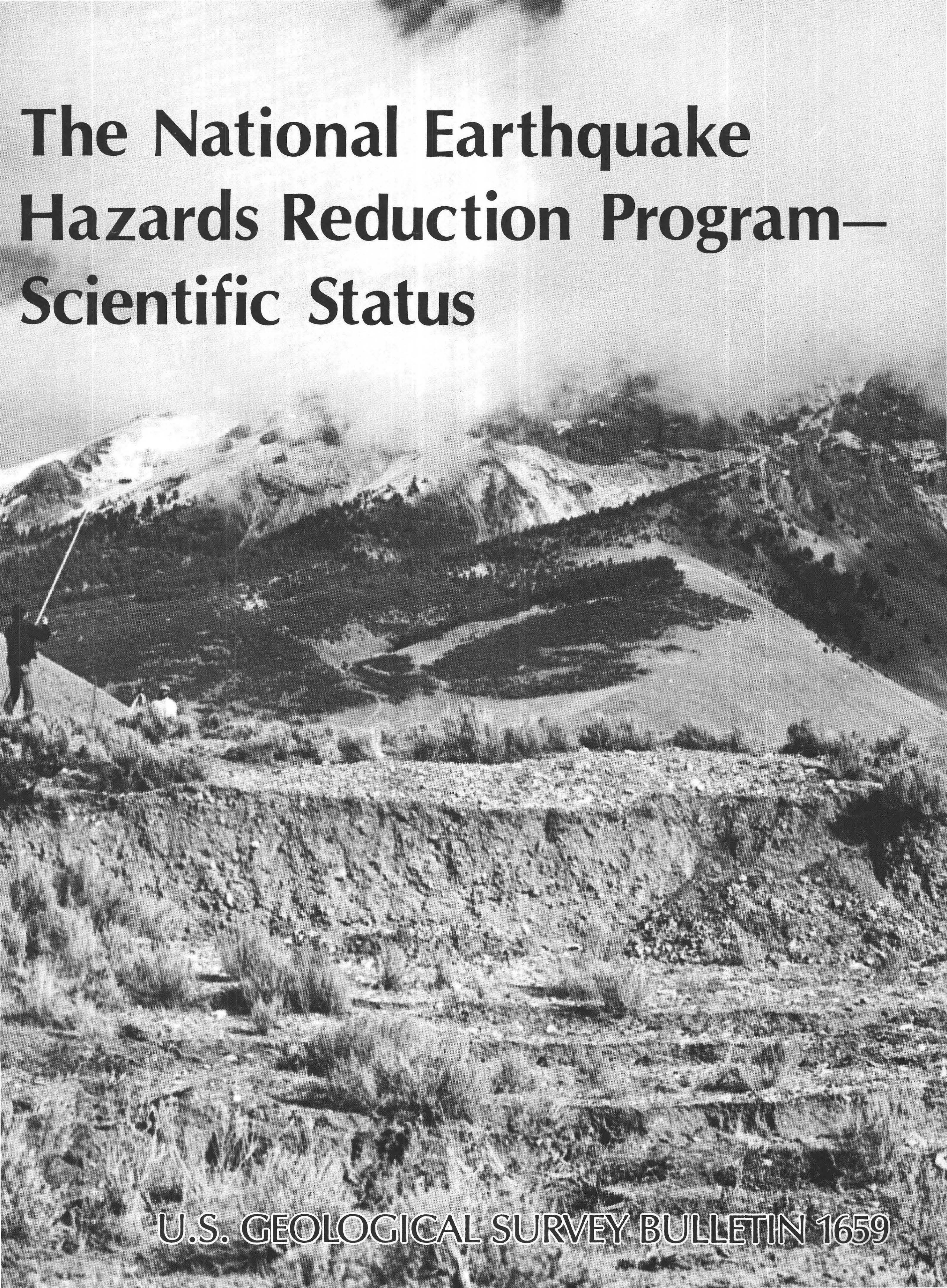
# The National Earthquake Hazards Reduction Program—Scientific Status



U.S. GEOLOGICAL SURVEY BULLETIN 1659

# The National Earthquake Hazards Reduction Program—Scientific Status

By THOMAS C. HANKS

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DEPARTMENT OF THE INTERIOR
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**COVER**

The main fault scarp of the **M** = 7.0 Borah Peak, Idaho, earthquake of October 28, 1983 (view is to the east, from a point just north of Doublespring Pass Road). The Lost River Range, in the background, rises to Borah Peak (elevation 12,655 ft), the highest point in Idaho, just to the right of the field of view. Over millions of years, the range has been uplifted to its present elevation and backtilted to the east along west-facing normal faults, such as the one which produced the scarp pictured here. The scarp cuts an alluvial fan of Pinedale age (12,000 to 15,000 years old). The beveled surface that intersects the 1983 scarp surface from above was, prior to the 1983 earthquake, the degraded fault scarp of a single earthquake that occurred several thousand years ago. Subsurface expression of this prehistoric earthquake was studied in detail on the walls of a trench excavated in 1976. This trench, located by the slightly darker and more erodable backfill material in the scarp face near the right margin of the cover, was displaced during the 1983 earthquake. Photograph courtesy of R. E. Wallace, U.S. Geological Survey.

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# The National Earthquake Hazards Reduction Program—Scientific Status

By Thomas C. Hanks

## INTRODUCTION

In the early morning of December 16, 1811, a major earthquake occurred near New Madrid, Missouri, just across the Mississippi River from the Kentucky-Tennessee state line. It was felt in Boston, some 1,800 km distant, and frightened people in Washington, D.C. In the same source area, comparable-size shocks occurred again on January 23 and yet again on February 7, 1812. Structural damage and loss of life resulting from these three earthquakes were slight, since the strongly shaken area was sparsely populated—then. Ground-failure effects, however, including sunken land, landslides, and liquefaction, occurred over an area of some 100,000 km$^2$. Navigation on the central part of the Mississippi River was uncertain for some time after the earthquakes, as a result of sunken islands, landslides and cave-ins, and uprooted timber. Only in the past decade has much progress been made in understanding the cause and effect of these earthquakes and the processes that will give rise to succeeding events and their attendant hazards in the New Madrid seismic zone.

At 0600 p.s.t. February 9, 1971, the northern Los Angeles Basin was strongly shaken by the San Fernando, California, earthquake. Although the collapse of two structures (built in 1925) at the San Fernando Veterans Hospital complex was responsible for most of the 58 people killed in the earthquake, modern construction, with only a few exceptions, fared very well—surely the result of earthquake-resistant design and construction procedures initiated in the Los Angeles area as a result of the 1933 Long Beach, California, earthquake. Only luck and a low water level, however, prevented the inundation of 80,000 people living in the flood plain beneath the lower Van Norman Reservoir; the dam itself failed, in any mechanical sense of the word (fig. 1). Even at an early date, an enormous amount of information on this earthquake and its effects was available, including a study of the psychological imprint the event left on small children awakened in the most intensely shaken areas.

Earthquakes have tormented the psyche and works of man from time immemorial, but nationally integrated and funded programs leading to meaningful earthquake-hazards reduction are only recent developments. In the United States, national interest in such a program began to develop during the early 1970's. The San Fernando earthquake itself was a major impetus, focusing attention again on earthquake hazards and the point that luck alone is no sure protection against them. At the same time, there was much scientific and technical reason to believe that meaningful reduction of earthquake hazards could be achieved through a national program: the scientific and engineering disciplines related to earthquakes and their effects had progressed rapidly through the 1960's, in part owing to the stimulus of the great 1964 Alaska earthquake.

A growing awareness that potentially damaging and destructive earthquakes are, indeed, a nationwide problem had also taken hold by the early 1970's. Figure 2 shows the locations of potentially damaging and destructive earthquakes that have occurred in the United States from colonial times through 1970. Although the "earthquake problem" is readily associated with Alaska, Hawaii, California, and Nevada, large earthquakes have also shaken parts or all of Washington, Oregon, Idaho, Montana, and Utah several times or more in the 20th century. Neither is the eastern seaboard any stranger to such earthquakes, although they occur less frequently here; in that region, the major seismic events of the historical record are the St. Lawrence River valley earthquakes of 1663 and 1925, the Cape Ann, Massachusetts, earthquake of 1755, and the Charleston, South

Carolina, earthquake of 1886. Geographic locations, dates, and magnitudes of all the earthquakes mentioned in the text are listed in table 1.

Two earthquakes in 1975 provided additional dimensions to the developing interest in a national program of earthquake-hazards mitigation. The first of these was the Haicheng earthquake in northeastern China and its spectacular prediction by Chinese scientists. Closely following promising reports of earthquake-prediction research in the Soviet Union, the prediction of the Haicheng earthquake provided much additional impetus for earthquake-prediction efforts in the United States. The second event was the Oroville, California, earthquake adjacent to the Oroville Dam, at the headwaters of the California Aqueduct. This earthquake reminded us anew that people can condition the occurrence of earthquakes by manipulation of water supplies, both through reservoir impoundment and through deep-level pumping for petroleum extraction or fluid-waste disposal.

These developments, together with the attendant public and political concern, culminated in the Earthquake Hazards Reduction Act of 1977 (Public Law 95–124). This act directed the President to establish a National Earthquake Hazards Reduction Program (NEHRP), with objectives as given in table 2. Public Law 95–124 authorized funding of NEHRP for three years; since fiscal year (FY) 1980, Congress has reauthorized NEHRP on an annual basis. As of FY 1984,

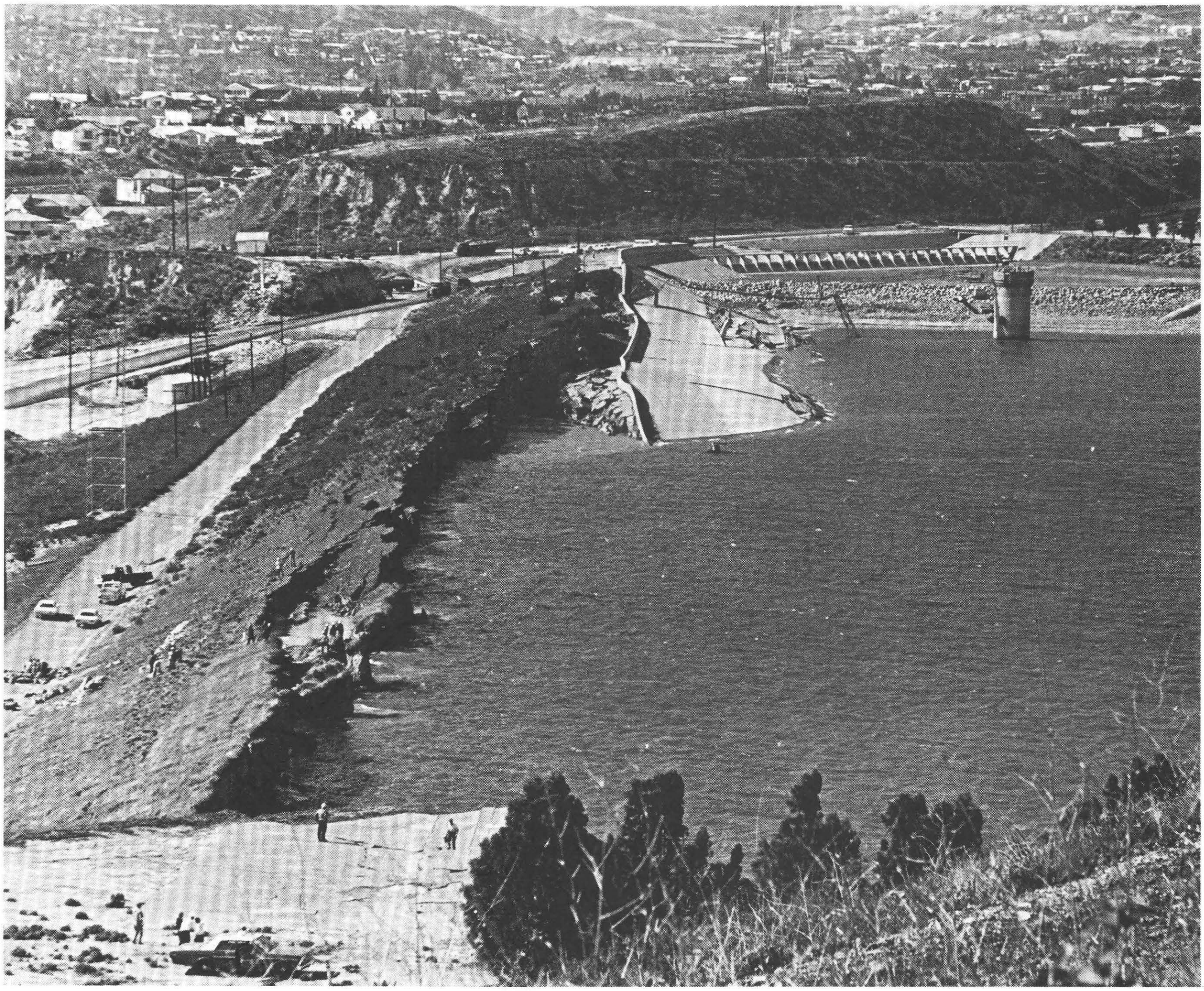


**Figure 1.** Lower Van Norman Dam (looking west) on the morning of February 9, 1971, several hours after the San Fernando earthquake. A massive slope failure of most of the upstream face has downdropped the damcrest into the reservoir, leaving just a few feet of freeboard on the remnant face. From R.F. Scott, "San Fernando Earthquake, 9 February 1971, Preliminary Soil Engineering Report," *in* Jennings, P.C., ed., "Engineering Features of the San Fernando Earthquake, February 9, 1971" (Pasadena, California Institute of Technology, Earthquake Engineering Research Laboratory Report EERL 71–02, p. 299–331), 1971.

four Federal agencies are funded directly through NEHRP: the U.S. Geological Survey (USGS; $35.3 million of which $8.4 million supports a grants and contracts program external to USGS), the National Science Foundation (NSF; $25.9 million), the Federal Emergency Management Agency (FEMA; $4.1 million), and the National Bureau of Standards (NBS; $0.5 million). Generally speaking, the USGS has responsibility for the scientific elements of the program, and the NSF has responsibility for the engineering elements, although the NSF Division of Earth Sciences supports basic research related to the NEHRP as well. The FEMA is charged with planning and coordinating the entire program, as well as implementing mitigation and preparedness strategies. The NBS has responsibility for developing building codes and for testing earthquake-resistant construction.

The purpose of this report is to examine the scientific status of NEHRP, that is, to describe and assess our present body of scientific understanding and data bases pertinent to the objectives of NEHRP. It is worth emphasizing, however, that the objectives of NEHRP cannot be met with scientific achievements alone. Such developments must sooner or later become part of rational implementation plans, land-use-planning policies, building codes, and emergency-preparedness plans, for example, which can do so much to reduce, if not obviate, potential hazards from earthquakes. We do not speak much to such implementation plans here, our point of view being the obvious but narrow one: these implementation plans may not mean much if they are not based on the best scientific knowledge and data available. Equally provincially, in speaking of "scientific understanding and data bases," we shall be concerned with those related to the earth sciences, not the engineering sciences. Although much of the scientific knowledge discussed herein is of great relevance to—and is routinely used in—problems of engineering concern, the reader will find little here about the many accomplishments of the earthquake engineering fields appropriate, say, to objective 1 in table 2.

## ACKNOWLEDGMENTS

This report has its origins in a set of letters in which the following scientists outlined their assessment of the significant scientific accomplishments related to the National Earthquake Hazards Reduction Program: C.R. Allen and H.

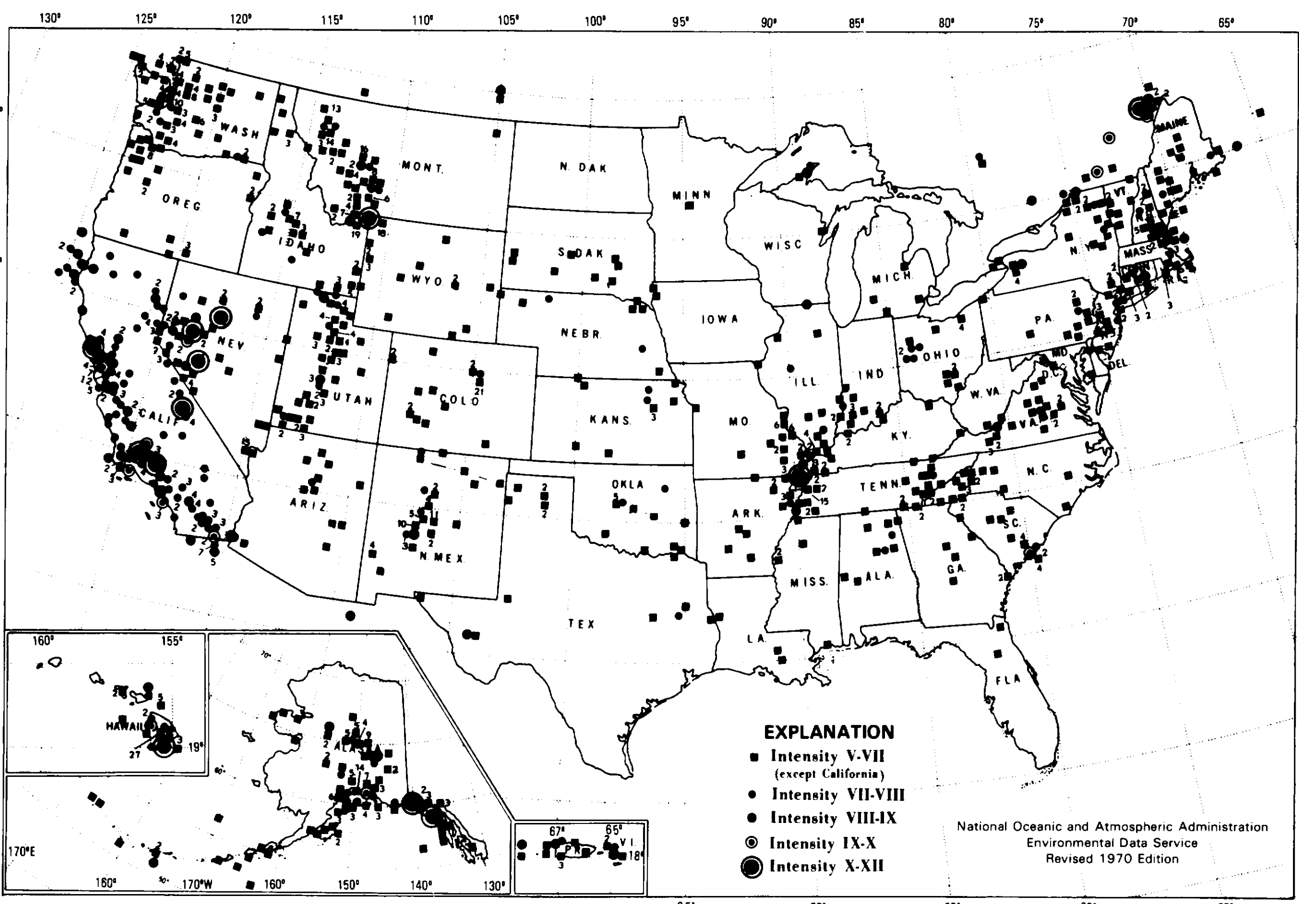



**Figure 2.** United States, showing locations of earthquakes of modified Mercalli intensity V or greater from colonial times through 1970. From "Earthquake History of the United States" (U.S. Department of Commerce, National Oceanic and Atmospheric Administration and U.S. Geological Survey Publication 41–1 revised edition (through 1970), reprinted 1982 with supplement (1971–80)).

## Table 1.

Locations, dates, and moment magnitudes of earthquakes mentioned in the text

| Location | Date | Magnitude |
|---|---|---|
| Honshu, Japan | Sept. 13, 1984 | 6.5 |
| Morgan Hill, California | Apr. 24, 1984 | 6.2 |
| Borah Peak, Idaho | Oct. 28, 1983 | 7.0 |
| Coalinga, California | May 2, 1983 | 6.4 |
| Gaza, New Hampshire | Jan. 18, 1982 | 4.5 |
| New Brunswick, Canada | Jan. 9, 1982 | 5.8 |
| Aswan, Egypt | Nov. 14, 1981 | 5.5 |
| Eureka, California | Nov. 8, 1980 | 7.3 |
| El Asnam, Algeria | Oct. 10, 1980 | 6.9 |
| Mammoth Lakes, California | May 27, 1980 | 6.0 |
| Mammoth Lakes, California | May 25, 1980 | 6.0 |
| Mammoth Lakes, California | May 25, 1980 | 5.7 |
| Mammoth Lakes, California | May 25, 1980 | 6.2 |
| Livermore Valley, California | Jan. 27, 1980 | 5.5 |
| Livermore Valley, California | Jan. 24, 1980 | 5.8 |
| Imperial Valley, California | Oct. 15, 1979 | 6.5 |
| Coyote Lake, California | Aug. 6, 1979 | 5.7 |
| Santa Barbara, California | Aug. 13, 1978 | 6.0 |
| Klerksdorp Mining District, South Africa | Apr. 7, 1977 | 5.5 |
| Oroville, California (aftershock) | Aug. 6, 1975 | 4.8 |
| Oroville, California | Aug. 1, 1975 | 5.8 |
| Haicheng, People's Republic of China | Feb. 4, 1975 | 7.0 |
| San Fernando, California | Feb. 9, 1971 | 6.6 |
| Koyna, India | Dec. 10, 1967 | 6.3 |
| Parkfield, California | June 28, 1966 | 6.1 |
| Kremasta, Greece | Feb. 5, 1966 | 6.3 |
| Prince William Sound, Alaska | Mar. 28, 1964 | 9.2 |
| Kariba, Zambia/Zimbabwe | Sept. 23, 1963 | 5.8 |
| Hsinfengkiang, People's Republic of China | Mar. 18, 1962 | 6.1 |
| Imperial Valley, California | May 19, 1940 | 7.0 |
| Parkfield, California | June 8, 1934 | 6.0 |
| Long Beach, California | Mar. 11, 1933 | 6.2 |
| St. Lawrence River Valley, Canada | Mar. 1, 1925 | 6.2 |
| Parkfield, California | Mar. 10, 1922 | 6.0 |
| San Francisco, California | Apr. 18, 1906 | 7.7 |
| Parkfield, California | Mar. 3, 1901 | $\simeq$6 |
| Charleston, South Carolina | Sept. 1, 1886 | 6.5–7 |
| Parkfield, California | Apr. 10, 1881 | $\simeq$6 |
| Owens Valley, California | Mar. 26, 1872 | 7.8 |
| Fort Tejon, California | Jan. 9, 1857 | 7.9 |
| Hayward, California | June 10, 1836 | 6.5–7 |
| New Madrid, Missouri | Feb. 7, 1812 | 7–7.5 |
| New Madrid, Missouri | Jan. 23, 1812 | 7–7.5 |
| New Madrid, Missouri | Dec. 16, 1811 | 7–7.5 |
| Cape Ann, Massachusetts | Nov. 18, 1755 | 6.5–7 |
| Ningxia-Hui Autonomous Region, People's Republic of China | Jan. 3, 1739 | 7.5–8 |
| St. Lawrence River Valley, Canada | Feb. 6, 1663 | 6–6.5 |

## Table 2.

Objectives of the National Earthquake Hazards Reduction Program, as written in Public Law 95–124

(1) The development of technologically and economically feasible design and construction methods and procedures to make new and existing structures, in areas of seismic risk, earthquake resistant, giving priority to the development of such methods and procedures for nuclear power generating plants, dams, hospitals, schools, public utilities, public safety structures, high-occupancy buildings, and other structures that are especially needed in time of disaster;

(2) The implementation, in all areas of high or moderate seismic risk, of a system (including personnel, technology, and procedures) for predicting damaging earthquakes and for identifying, evaluating, and accurately characterizing seismic hazards;

(3) The development, publication, and promotion, in conjunction with State and local officials and professional organizations, of model codes and other means to coordinate information about seismic risk with land-use policy decisions and building activity;

(4) The development, in areas of seismic risk, of improved understanding of, and capability with respect to, earthquake-related issues, including methods of controlling the risks from earthquakes, planning to prevent such risks, disseminating warnings of earthquakes, organizing emergency services, and planning for reconstruction and redevelopment after an earthquake;

(5) The education of the public, including State and local officials, as to earthquake phenomena, the identification of locations and structures which are especially susceptible to earthquake damage, ways to reduce the adverse consequences of an earthquake, and related matters;

(6) The development of research on—
   (A) ways to increase the use of existing scientific and engineering knowledge to mitigate earthquake hazards;
   (B) the social, economic, legal, and political consequences of earthquake prediction; and
   (C) ways to assure the availability of earthquake insurance or some functional substitute; and

(7) The development of basic and applied research leading to a better understanding of the control or alteration of seismic phenomena.

Kanamori, California Institute of Technology; C.H. Scholz, D.W. Simpson, and L.R. Sykes, Columbia University; R.K. McGuire, Risk Engineering, Inc.; M.D. Zoback, Stanford University; R.D. Borcherdt, R.C. Bucknam, W.L. Ellsworth, E.R. Engdahl, J.F. Evernden, T.C. Hanks, D.P. Hill, K.R. Lajoie, A. McGarr, S.A. Sipkin, W. Thatcher, R.E. Wallace, and R.L. Wesson, U.S. Geological Survey; T.V. McEvilly, University of California, Berkeley; M. Wyss, University of Colorado; and K. Aki, University of Southern California. These letters were the basis of detailed discussions, roughly along the lines of the major sections of this report, held March 20, 21, and 22, 1983, at Asilomar, California. Attending this meeting were Borcherdt, Bucknam, Ellsworth, Engdahl, Evernden, Hanks, Hill, Kanamori, McGarr, McGuire, Simpson, Sipkin, Thatcher, Wallace, Wyss, and Zoback, as did J.F. Davis, California Division of Mines and Geology. On the basis of these discussions, these scientists wrote detailed contributions to each of the principal sections of this report. Writing from both the initial letters and the detailed Asilomar statements, Hanks compiled this report, with the additional help of M.L. Zoback's thoughts on the Basin and Range province. Carol Sullivan attended to the innumerable minutiae that accompany the formulation, preparation, and redaction of a manuscript such as this one.

## EARTHQUAKE STUDIES AND NEHRP

The essence of earthquake studies is their observational and empirical character. We know what we know about earthquakes almost entirely through what we observe about them or their laboratory analogs, not because we have any special insight into how the Earth works in the ways that generate earthquakes. Matters are likely to stay this way for the foreseeable future, since the processes that give rise to earthquakes and the material failure that results in them are extraordinarily complicated systems. Moreover, these are three-dimensional systems we must study remotely from a two-dimensional surface, and these processes operate on time scales long compared to a human lifetime.

Consistent with these circumstances are several other matters that strongly condition the ways in which progress is made in earthquake studies. In the first place, the great complexity of the cause and effect of earthquakes requires that a wide spectrum of earth science and engineering disciplines be brought to bear on the problem. Secondly, systematic, long-term observational programs—for example, those spanning two or more large earthquakes along the same fault—are of the utmost importance. Finally, because earthquakes know no political boundaries, there is much to be gained from continuing, shared interest in and cooperation among the many domestic, foreign, and international programs in earthquake studies.

With these considerations in mind, it will come as no surprise at all that a primary focus of NEHRP has been the systematic and comprehensive collection of fundamental data sets. These observations range over a wide spectrum of geological, geochemical and geophysical disciplines. They include, but are not limited to, geologic mapping on regional, local, and ~1-meter scales; assemblages of late-Quaternary dates and rates of deformation; global, regional, and local seismic monitoring of earthquakes; measurements of modern vertical and horizontal crustal deformation across wide ranges in spatial and temporal sampling; determination of regional crustal structure from seismic and gravity data; measurements of heat flow and *in situ* shear stress in tectonically active areas; recording of earthquake strong ground motions; and documentation of the effects of earthquakes, including surface faulting, dynamically induced ground failure (liquefaction and landslides), and diverse kinds of structural damage.

In most cases, however, these data sets—and NEHRP—are what they are today because of long histories of shared interests. The recording of earthquake strong ground motion, for example, stands at the interface between seismology and earthquake engineering. These observations are used by seismologists to elucidate dynamic characteristics of the faulting process, but also by engineers to understand the dynamic responses of both geologic and engineering structures; in many cases, the same record has served both purposes. Since NEHRP has been in place, several significant strong-motion-data sets have been collected in California alone, including those for the earthquakes of Santa Barbara (1978), Coyote Lake (1979), Imperial Valley (1979), Livermore Valley (1980), Mammoth Lakes (1980), Coalinga (1983), and Morgan Hill (1984). Certainly, the NEHRP-funded strong motion program of the USGS has figured prominently in the collection, processing, and analysis of these observations—but by no means exclusively. In California, the U.S. Army Corps of Engineers, the U.S. Bureau of Reclamation, the California Division of Mines and Geology, the Los Angeles Metropolitan Water District, and the U.S. Veterans Administration, as well as the USGS, all have significant stakes in the present state-wide configuration of strong-motion accelerographs, as do various universities operating special-purpose research arrays (generally funded through the NSF) and the sum total of building owners required to have strong-motion instrumentation by the Los Angeles Building Code and, more recently, by the Uniform Building Code. Moreover, the present-day strong-motion program of the USGS has antecedents dating back to 1932. Such continuing and common interests of diverse disciplines in various observational programs are the norm, not the exception, in earthquake studies—and in NEHRP.

What this means, then, is that NEHRP cannot be sharply circumscribed as a scientific entity in and of itself. NEHRP relies heavily in a temporal sense on the rich history of past observational experience and in a spatial sense on the many investigations carried on at numerous institutions, both at home and abroad. In the United States, many such studies are supported directly by NEHRP; from an international point of view, most of them are not. In this report, which purports to assess the scientific status of NEHRP as of 1984, we have borrowed freely from this larger experience of earthquake studies, although we cannot recount it entirely here. We certainly do not wish to imply that the scientific results described herein are due to NEHRP alone. Our predilections, indeed, are in just the opposite direction. We regard the continuity of observational traditions and of close cooperation with other programs and institutions as, if not exactly "accomplishments" of NEHRP, certainly essential features of its well-being. Simply stated, then, this report attempts to describe what "we" (in a large and shared sense) have learned scientifically since the inception of NEHRP—in terms that we know are generalized but that we hope are manageable—about these problems of national and international concern.

## TRENDS OF THE SCIENCE

At the heart of this science lie two basic questions: why do earthquakes occur, and what happens when they do? The latter question is associated with the gamut of earthquake hazards, from surface faulting, through strong ground motion, to ground and structural failures. The "why" of earthquakes can be reformulated as a set of questions related to where, when, how big, and by what mechanism, questions intimately connected with earthquake prediction, although they pertain to induced seismicity and earthquake control as well.

Whatever their association with earthquake hazards, prediction, and control may be, these questions set long-term themes for earthquake research. In this section, we discuss four such themes, trends of the science in the 1980's that have already yielded broad and general insights concerning the cause and effect of earthquakes. These matters, in their general importance, sit somewhat above the 10 more specific accomplishments presented in the following section. More importantly, these trends hold much promise for providing the science with an even more complete and general understanding of earthquakes.

### The measure of an earthquake

Most of what we know about earthquakes, specifically their locations (including origin times), faulting orientations, source strengths, and dynamic faulting characteristics, is derived from seismic data. Such observations tell us that an earthquake occurred, when and where it occurred, and, as a result of quite recent developments, increasingly more about how it occurred. Seismic data alone, however, do not tell us *why* a certain earthquake occurred, nor do they tell us when the *next* such earthquake will occur; to address such questions, we must turn to a host of other geologic and geophysical observations. Nevertheless, earthquake catalogs that are accurate and complete with respect to the above characteristics are essential to earthquake studies because earthquake-recurrence intervals, seismic-moment-release rates (or fault-slip rates), strong ground motion, and earthquake risk, among other things, are assembled from such data.

By long custom, earthquake locations, magnitudes, and faulting orientations have been the principal data sets of earthquake seismology. After the installation of the World-Wide Standardized Seismological Network (WWSSN) in the early 1960's, such determinations became greatly improved on a global basis and played a central role in the development of the theory of plate tectonics and in matters relating to the discrimination of earthquakes from underground nuclear explosions. Since the early 1970's, rapid progress has been made in the development of broad-band, high-dynamic-range, digital instrumentation. Global, regional, and local deployment of arrays of such instruments are now providing data that are significant in many important ways. In this section, we describe how such data have been used to take the measure of an earthquake. In other sections, we describe results that depend heavily on these same advances, including fully computerized locations of the swarm of earthquakes at Mammoth Lakes, California, in January-February, 1983, determination of the crustal structure of the New Madrid seismic zone, and waveform inversion for the source mechanism of the 1979 Imperial Valley earthquake.

(It is impossible to overstate the significance of these advances in digital data acquisition and in computer technology allowing for rapid and efficient digital data processing. No other science has the data problems that earthquake seismology has: from the smallest earthquake we can measure under normal circumstances to the largest earthquake we know of, signal amplitudes span a factor of $10^{10}$, across a frequency band of six orders of magnitude or more. These systems, moreover, must always be at the ready, so to record a partial set of the hundreds of thousands of earthquakes that are registered globally each year.)

Seismic moment and the seismic moment tensor (which specifies the directional nature of faulting in terms of the slip vector and fault-surface orientations) are readily, accurately, and now routinely determined from such digital data (fig. 3). Seismic moment, $M_0 = \mu \bar{u} A$ where $\mu$ is the shear modulus of the source region and $\bar{u}$ is the average slip across the fault surface of area $A$, figures prominently in earthquake studies. Cumulative moment sums are easily translatable into average slip rates for any fault of interest, and nowadays it is almost meaningless to speak of a recurrence interval without specifying the $M_0$ of the event of interest. Seismic moment, moreover, is the controlling source factor in strong-ground-motion excitation, inasmuch as the dynamic-stress differences that accompany crustal faulting appear to be so remarkably constant (see discussion below). The most important practical value of rapid and accurate determinations of the seismic-moment tensor, however, is for tsunami warning, in the case of major shallow earthquakes that rupture the ocean floor. Including the transit times of the seismic waves, seismic-moment-tensor inversions, in principle, can be available within tens of minutes of the earthquake occurrence, whereas tsunami-transit times are hours or more to distant landfalls.

It is moreover known that $M_0$ does not suffer from the "saturation" problems that afflict all time-domain-amplitude magnitude scales. The recent development of a moment magnitude scale, $\mathbf{M} = \frac{2}{3} \log M_0 - 10.7$, is a natural progression to a more quantitative, more accurate, and more physically meaningful statement of source strength. Several recent studies have shown that $\mathbf{M}$ agrees well with $M_L$ (local magnitude) for $3 \lesssim M_L \lesssim 7$, with $M_s$ (surface-wave magnitude) for $5 \lesssim M_s \lesssim 7\frac{1}{2}$, and with $M_w$ (energy magnitude) at larger magnitudes. The moment-magnitude scale, then, promises to provide earthquake studies with a unifying and uniformly valid measure of source strength.

In addition to seismic moment, other measures of the earthquake source may be extracted from seismograms, including faulting duration, dimension and direction of rupture, various stress differences accompanying the failure process, and measures of the nature and extent of dynamically inhomogeneous faulting. In the past decade, much interest has been focused on the stress differences associated with crustal faulting, in part for what this tells us about the failure process itself and the conditions that drive it and in part for the strong influence such stress differences exert on the resulting ground motion.

It has been known for some time that the average static-stress drops of earthquakes (that is, the difference between stress levels before and after faulting) are essentially constant over the entire range of events accessible by instrumental recording ($10^{16} \lesssim M_0 \leqslant 10^{30}$ dyne-cm; $0 \lesssim \mathbf{M} \lesssim 9.5$). Typically, these stress drops range from several bars to several hundred bars, and although this variance is not small on an absolute scale, it is indeed small in comparison with the range of seismic moments and source dimensions of earthquakes across the entire range $0 \lesssim \mathbf{M} \leqslant 9.5$. The source of this "stress-drop variance" is not well understood, although part of it is attributable to true uncertainties in the static-stress-drop determinations. At least some of this variance is thought to be of true tectonic significance, but it has been noteworthy in this regard that the stress drops of tremors induced by deep gold mining in South Africa are indistinguishable from those of naturally occurring, tectonically driven earthquakes.

In the past decade, several estimates of the dynamic-stress differences accompanying crustal faulting have become available, all of them obtainable from some measure of the recorded ground acceleration. The significance of these results are several. At any given depth, the dynamic-stress differences show less scatter than do the average static-stress drops. Second, very recent studies of dynamic-stress differences in California and Japan reveal that they depend on both depth and faulting type—dependences not discernible in static-stress-drop estimates, probably because of their larger uncertainties. These results point to a frictional model of faulting wherein dynamic stress differences should increase with depth in the seismogenic layer. Finally, because various measures of ground acceleration of engineering interest are directly controlled by these dynamic-stress differences, an important practical result of this approach has been the development of simple but accurate techniques to calculate ground-motion amplitudes for engineering design purposes, a matter to which we return in the section below entitled "Strong Ground Motion."

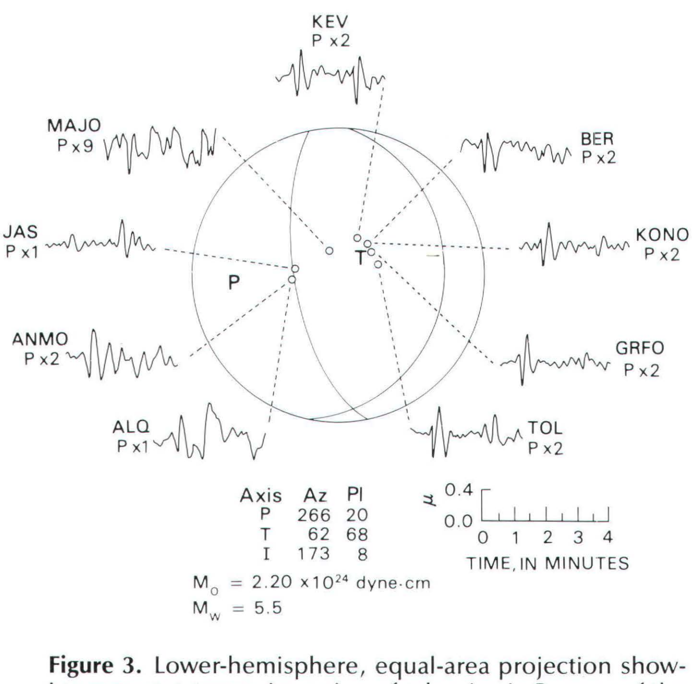



**Figure 3.** Lower-hemisphere, equal-area projection showing moment-tensor inversion of teleseismic *P* waves of the 1982 New Brunswick, Canada, earthquake. Initial analysis from U.S. Geological Survey, Branch of Global Seismology and Geomagnetism.

## The earthquake cycle

The notion of an earthquake cycle dates back to the Reid hypothesis, formulated by H.F. Reid on the basis of his observations of the 1906 San Francisco earthquake. This hypothesis holds that two comparable-size earthquakes rupturing the same section of fault will be separated by a period of time sufficient to reaccumulate strain by an amount equal to the elastic-strain drop accompanying the first earthquake. Since then, specific stages within the earthquake cycle have been recognized for major earthquakes: a long period of seismic quiescence following a major earthquake and its immediate aftershocks, a shorter (and varying) period of enhanced seismicity as elastic-strain accumulation approaches (and, locally, exceeds) the critical strain level, and a very short (hours to days) but commonly nonexistent period of immediate foreshocks, all followed by the next major earthquake.

Segments of plate boundaries that have exhibited long periods (decades to centuries) of seismic quiescence are now known as seismic gaps. They can be mapped worldwide, and systematic identification of them has proceeded through the 1970's. Initially, seismic gaps were ranked in terms of their "seismic potential" solely on the basis of the length of time since the last, gap-filling earthquake. Between 1968 and 1980, 10 major ($\mathbf{M} \gtrsim 7.5$) plate-boundary earthquakes were correctly anticipated by means of the seismic-gap hypothesis (fig. 4).

This success in forecasting the place and magnitude of major plate-boundary earthquakes is of great importance in rationally anticipating, planning for, and working against potential earthquake hazards. The concept is also important to the siting and execution of experimental programs with which we may determine more precisely the time of the impending earthquake, as well as to the siting of strong-motion accelerographs to obtain ground-motion records close to major and great earthquakes. In the United States, the Shumagin Islands, Alaska, seismic gap and the section of the San Andreas fault that broke in the 1857 Fort Tejon, California, earthquake are considered to be the most likely sites for the next $\mathbf{M} \cong 8$ earthquakes (fig. 4). More recently, the essential features of the earthquake-cycle concept have been formalized in the time-predictable model of earthquake occurrence. The essence of this model is the same as the Reid hypothesis: the time to the next large earthquake can be estimated from the strain drop (or faulting displacement) of the preceding large earthquake and the rate at which strain (or displacement) is accumulating. For six situations in which three or more earthquakes are known in the historical record, predicted repeat times are in agreement with the observed repeat times to an accuracy of 2 to 50 percent.

An important ingredient of the earthquake-cycle model, which in fact is an assumption of the Reid hypothesis, is that all of the strain accumulated during the interevent period will be released seismically—that is to say, all of the strain accumulation is elastic. For many of the major fault systems of the world, this seems to be the case: elastic-slip rates obtained from seismic-moment sums are in reasonable agreement with long-term (millions of years) fault-slip rates, as obtained from geologic or paleomagnetic data—a result in keeping with the tenets of rigid-plate tectonics. Although some uncertainty is attendant to this coincidence owing to the less reliable seismic data of the early 1900's and before, and although the significance of measurable permanent (nonelastic) tectonic deformation has only recently been appreciated, the importance of continually monitoring strain accumulation in seismically active areas can hardly be overestimated.

In fact, a variety of studies can help to determine what stage of the earthquake cycle some particular fault segment is in, at least for continental areas. Studies of paleoseismicity, especially those involving detailed investigations of the

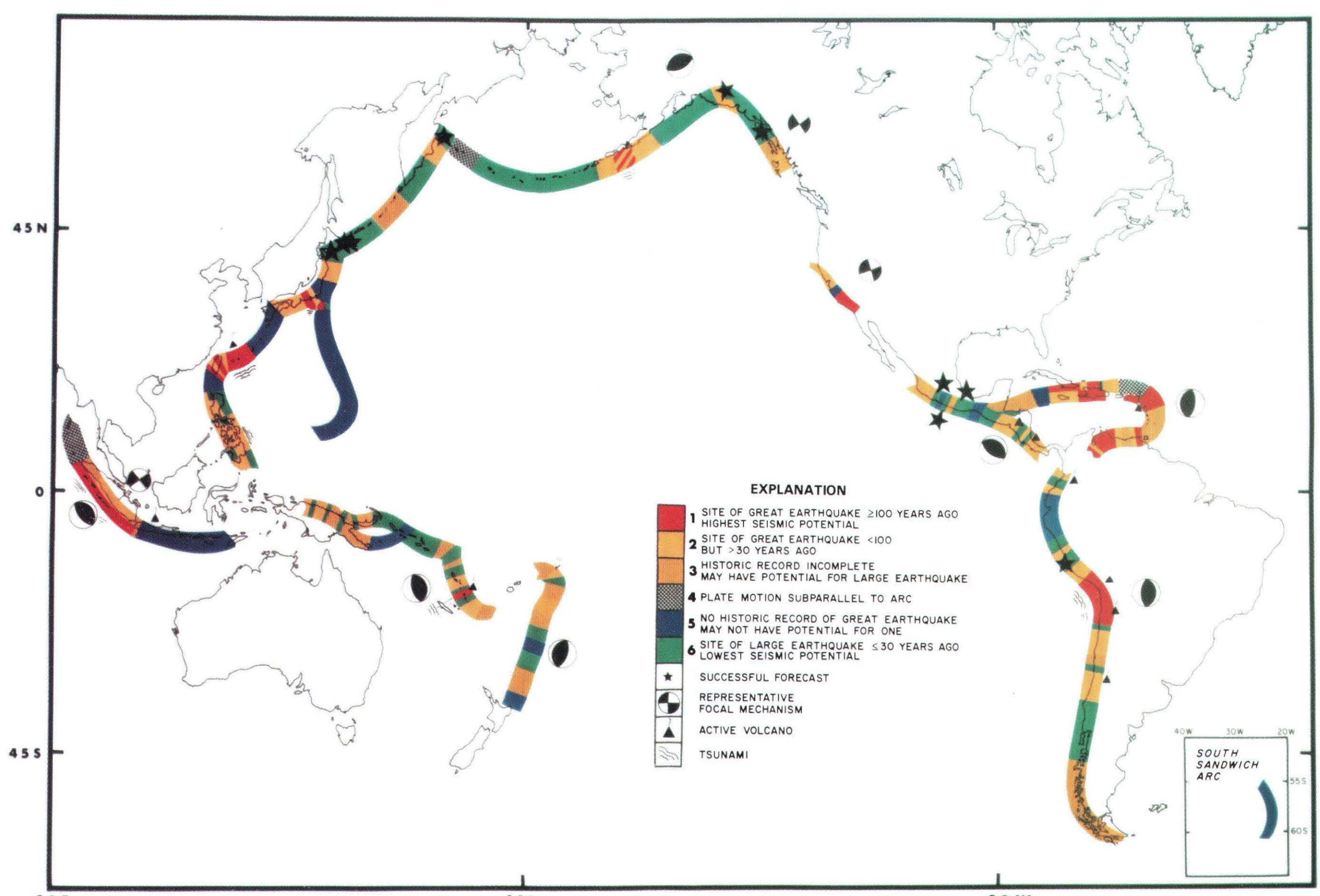



**Figure 4.** Seismic potential of the major circum-Pacific plate boundaries, as reckoned in 1980. Updated from W.R. McCann, S.P. Nishenko, L.R. Sykes, and J. Krause, "Seismic Gaps and Plate Tectonics: Seismic Potential for Major Plate Boundaries" (Pure and Applied Geophysics, v. 117, p. 1082–1147), 1979.

Holocene history of faulting, are of great value in estimating average fault-slip rates and recurrence intervals between major earthquakes. Seismicity studies can reveal any temporal variation in earthquake occurrence related to an impending large earthquake. Advances in these specific study areas, as well in crustal-deformation studies, are detailed in later sections of this report.

The results of such investigations have recently been synthesized to estimate the time of occurrence of the next great San Francisco earthquake, reckoned as a "repeat" of the 1906 earthquake. On the basis of *average* slip rates and strain accumulation since 1906, the repeat of such an earthquake appears to be 50 years or more into the future. On the other hand, the recent increase of seismicity at the $\mathbf{M} \geqslant 5$ level in and around the San Francisco Bay region (fig. 5) suggests that the $6 \lesssim \mathbf{M} \lesssim 7$ earthquakes that preceded the 1906 earthquake may return to the bay region at any time.

These are promising developments with respect to the long-term prediction of major earthquakes. Continuing development of these ideas, however, can only proceed at the rate at which new observations are collected—which is slow in the case of major plate-boundary earthquakes. With no less concern for the major gaps of highest seismic potential (fig. 4), much interest is now directed to smaller fault segments, where $\mathbf{M} \cong 6$ earthquakes occur more frequently. Near Parkfield, California, at the northwest extremity of the 1857 rupture of the San Andreas fault, comparable-size ($5\frac{1}{2} \lesssim \mathbf{M} \lesssim 6$) earthquakes have occurred with a striking regularity (1881, 1901, 1922, 1934, and 1966). It seems quite likely that another such event will occur within the next 5 years. Some concern exists that such an event might trigger a repeat of the great 1857 earthquake, because the 1857 earthquake is inferred to have initiated near Parkfield and because the time elapsed from the 1857 event is now close to the average repeat time for such events as revealed at Pallett Creek (see fig. 13 and attendant discussion).

## Structural and Tectonic Setting

Underlying all considerations of the structural and tectonic setting of earthquakes is the unifying theme of plate tectonics. Most of the world's earthquakes and nearly all great ($\mathbf{M} \geqslant 8$) earthquakes occur along the major (nonridge) plate boundaries (fig. 6). Most of the results of the preceding section, in fact, can be regarded as natural corollaries of rigid-plate tectonics. These ideas do not, however, explain all aspects of crustal earthquakes; in fact, rigid-plate tectonics fails to explain most crustal earthquakes in continental areas—just those earthquakes of greatest social concern. In the United States, the distribution of seismicity and faults in southern California is left unexplained by rigid-plate tectonics, as is the seismicity and faulting throughout the Basin and Range province and in the eastern and central United States as well. Neither do the recent 1983 earthquakes at Coalinga, California, and Borah Peak, Idaho, have much to do with these ideas. Volcanic and induced seismicity, of course, are also separate issues.

In all of the situations described immediately above, an understanding of the structural and tectonic setting beyond that provided by rigid plate tectonics is necessary to explain why these earthquakes occurred, when they might occur again, and where else such relatively infrequent events might occur. Several of these situations are treated separately in the following section, in particular, those advances related to the understanding of earthquakes in the eastern and central United States, volcanic seismicity, and induced seismicity.

For the western United States, a scaled-down version of plate tectonics, "contemporary block tectonics" has proved useful in understanding the structural kinematics and relative motions of crustal blocks adjacent to the plate boundary in California and Nevada. Figure 7 illustrates details of the concept for the southern California region. An important result is that the Hosgri fault, a major right-lateral strike-slip

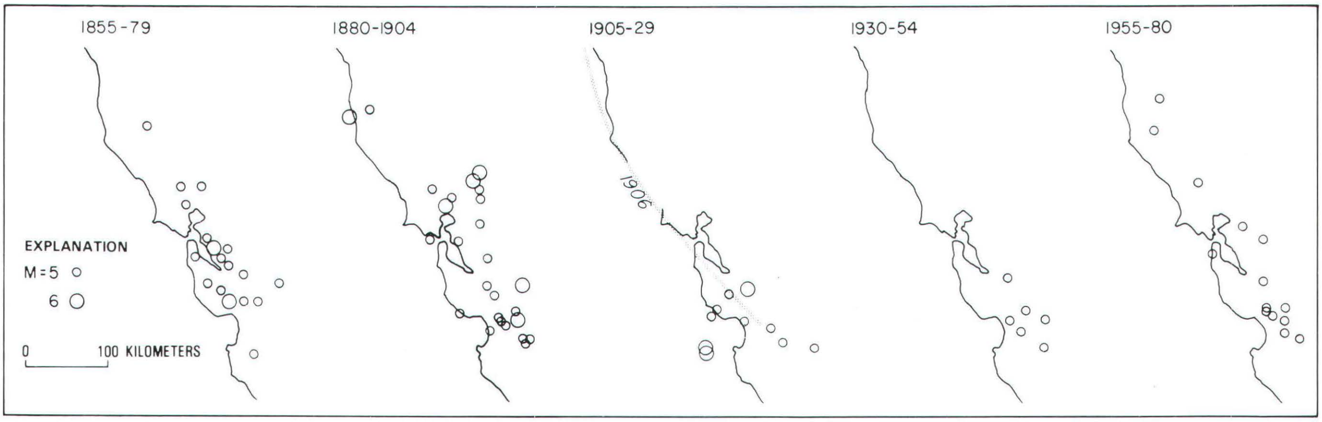



**Figure 5.** Sketch maps showing seismicity of the San Francisco Bay region ($\mathbf{M} \geqslant 5$) in quarter-century intervals before and after the 1906 San Francisco earthquake, which ruptured the San Andreas fault along the shaded zone in the middle frame. From W.L. Ellsworth, A.G. Lindh, W.H. Prescott, and D.G. Herd, "The 1906 San Francisco Earthquake and the Seismic Cycle," *in* Simpson, D.W., and Richards, P.G., eds., "Earthquake Prediction: An International Review" (Washington, American Geophysical Union, p. 126–140), 1981.

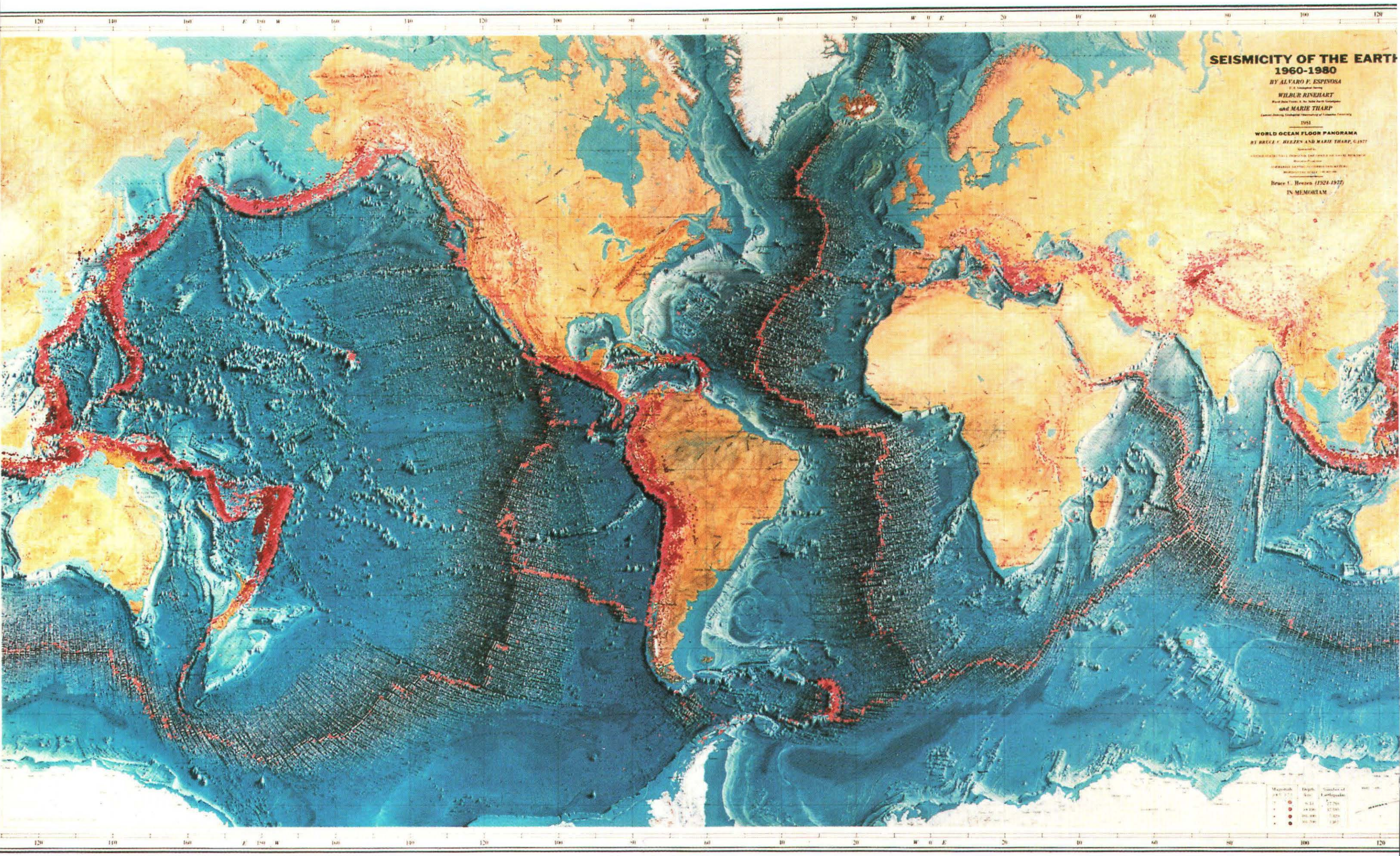



**Figure 6.** Seismicity of the Earth, 1960-80. Compiled by A.F. Espinosa, W. Rinehart, and M. Tharp on the "World Ocean Floor Panorama" of B.C. Heezen and M. Tharp, 1982. Reprinted with permission.

fault subparallel to and offshore of the San Andreas fault in central California, must undergo a component of east-west compression, and this prediction is borne out by focal mechanism studies. Just as in rigid-plate tectonics, block tectonics holds that seismicity should be concentrated at block boundaries, with little or no deformation internal to the block.

A parallel to these ideas exists in the finding that much of the seismicity in the eastern United States is apparently controlled by ancient zones of weakness, reactivated by the modern stress regime. The possibility that these weak zones define modern block boundaries could be of much importance in defining "seismic source zones" in the eastern United States, even if this is not the only reason for earthquakes in the region.

Yet a higher level of detail in how the structural and tectonic setting conditions earthquake occurrence may be sought in the synthesis of various kinds of geologic and geophysical data. In a later section, we describe how such information has led to a greatly improved understanding of the workings of the New Madrid, Missouri, seismic zone. Just below, earthquake occurrence in the larger scale Basin and Range province is considered in the light of such developments.

For the past 30 million years or so, the region between the Wasatch front in central Utah and the Sierra front in eastern California has opened by as much as several hundred kilometers. The extensional tectonics of the Basin and Range province, which may accommodate as much as 25 percent of the relative motion between the North American and Pacific plates in these latitudes, is evidenced by all manner of geologic, geophysical, and seismologic phenomena, including a thin crust, high heat flow, extensive magmatism and hydrothermal mineralization, low seismic velocities, high attenuation of seismic waves, and local compensation of individual basin-and-range pairs. This extensional deformation continues to the present, as evidenced by modern geodetic measurements (see fig. 11) and the normal faulting associated with historical earthquakes. Hydrofracture measurements have, in addition, revealed that the Basin and Range province is only marginally stable with respect to normal faulting, according to frictional-failure models.

In the brittle, seismogenic upper crust, the WNW-ESE extension of the Basin and Range province is accommodated by normal faulting, which produces northerly striking, alternating sequences of uplifted and backtilted mountain ranges and downdropped, sediment-filled basins, from which the

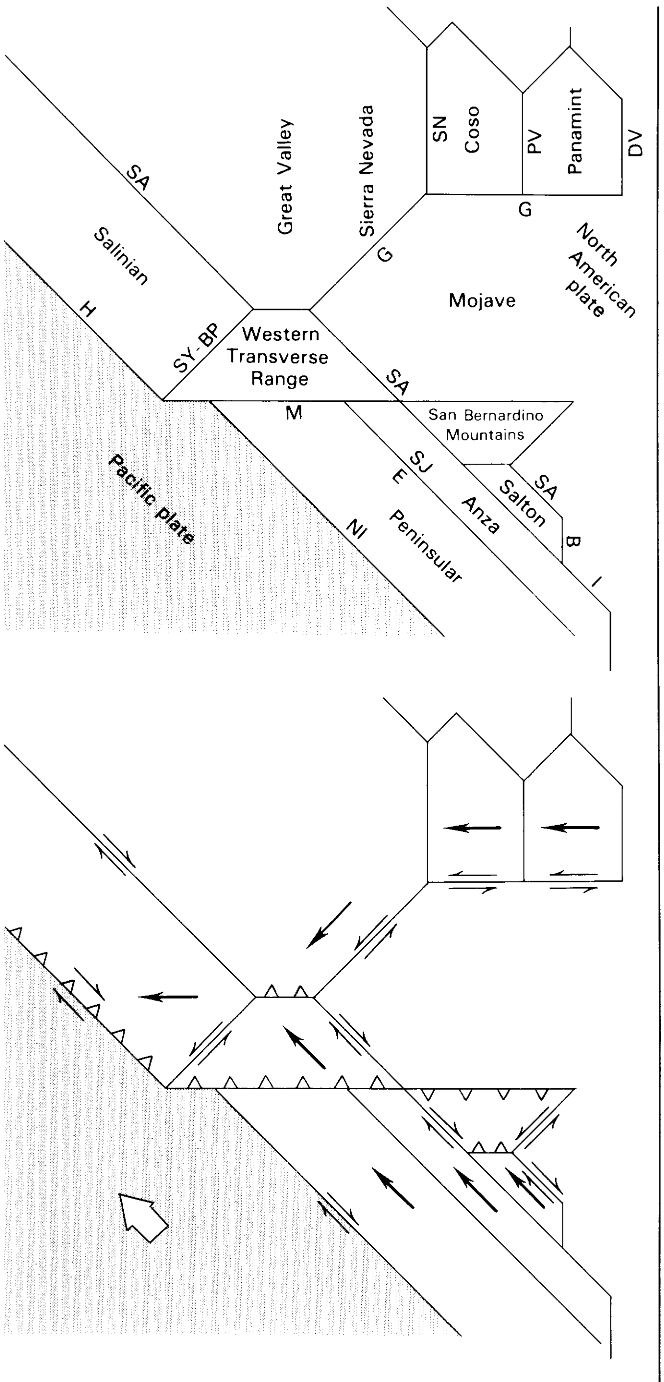



**Figure 7.** Idealized tectonic blocks of southern California. *Top,* initial configuration. Faults: **B,** Brawley; **DV,** Death Valley; **E,** Elsinore; **G,** Garlock; **H,** Hosgri; **I,** Imperial; **M,** Malibu; **NI,** Newport-Inglewood; **PV,** Panamint Valley, **SA,** San Andreas; **SJ,** San Jacinto; **SN,** Sierra Nevada; **SY-BP,** Santa Ynez-Big Pine. *Bottom,* block motion following incremental displacement in response to north-south compression and east-west extension. From D.P. Hill, "Contemporary Block Tectonics: California and Nevada" (Journal of Geophysical Research, v. 87, p. 5433–5450), 1982.

province takes its name. The approximately uniform spacing of the basin-and-range pairs (typically, 25–35 km) is clear evidence that the extension across the province has been fairly uniform, at least on a time scale of millions of years. For hundreds of thousands of years and less, however, this deformation has been decidedly nonuniform (fig. 8), and this poses important problems not only for the mechanics of this extensional deformation but also for the realistic assessment of earthquake hazards in this region. The major earthquakes of the historical record, for example, have occurred entirely within a narrow corridor in the west-central part of the province (fig. 8). Nevertheless, there is clear evidence that the Wasatch front has been faulted numerous times by $\mathbf{M} \cong 7$ earthquakes in the past several thousand years (fig. 9), and a scarp along the North Creek section of the Wasatch fault zone is so fresh that it is unlikely to be older than several hundred years.

Recent studies have focused on the subsurface geometry of faults and faulting and the mechanisms of deep-level extension. Analysis of a growing collection of seismic-reflection data and borehole logs has revealed a complex hierarchy of faults responsible for the modern deformation. In some areas, low-angle normal faults have been detected at only 4- to 6-km depth, and these apparently serve as detachment surfaces, because high-angle normal faults observed at the Earth's surface merge with (but do not cut) these detachment surfaces. In other places, however, these high-angle normal faults evidently penetrate to depths of 10 to 15 km. These results point to varying mechanisms and sizes of earthquakes, depending on the nature and extent of the detachment surfaces and the mechanism by which they accommodate extensional displacements.

## Inhomogeneous faulting

Heterogeneous structures along active crustal fault zones apparently exist on all length scales, as do the earthquakes that occur along them, from wavelengths of a thousand kilometers or so down to microscopic grain-size dimensions. It has been shown many times that the size of an earthquake is controlled by offsets or discontinuities separating otherwise-coherent fault segments, again across a wide range in length scales. Many of the major gap-filling earthquakes, for example, have filled the entire preexisting seismic gap, the end points of which are commonly associated with clear evidence for structural discontinuity. On the San Andreas fault, on a somewhat smaller scale, the southern extent of the 1906 earthquake and the northern extent of the 1857 earthquake (and future recurrences of either) are probably controlled—and isolated from each other—by the 180 km of the "creeping" section of the fault that lies in between. On yet a smaller scale, the Wasatch fault in central Utah and southern Idaho apparently breaks in earthquakes the size of which are controlled by structural discontinuities along strike (fig. 9). On scales of just several kilometers or so, normal faults in the Albuquerque-Belen and La Jencia Basins within the Rio Grande rift valley of central New Mexico demonstrate discontinuous behavior spatially and highly sporadic movement temporally. On an even smaller scale, a ½-km offset of the trace of the San Andreas fault near Parkfield is associated with clear mechanical complexity at depth, which includes an offset of the plane of seismicity at depth and fault-plane solutions significantly rotated from the trend of the fault. This offset is now thought to be the southeast terminus of the 1966 Parkfield earthquake. Seismicity studies in the 1980's now allow us to examine such heterogeneous structures throughout the seismogenic depth on even smaller length scales.

From such observations over such wide ranges in scale, it is apparent that inhomogeneous fault structures play two important roles in generating earthquakes. In the first place, one may regard the faults themselves as inhomogeneities, separating the stronger, less deformable plates or blocks on either side. In this role, the inhomogeneity concentrates deformation to the faults themselves; the ideas of plate or block tectonics follow naturally from this mechanism. Second, inhomogeneous structures along or within fault zones control the size or length of earthquakes along any particular fault. This has led to the recent development of the characteristic earthquake model, the idea that any particular fault segment will tend to rupture in a certain size event, the length of that event being controlled by binding structures along the fault. The rupture of these high-strength structures, in turn, can heavily condition the spectral composition of the resulting ground motion.

The significance of the characteristic-earthquake model is that it allows an estimate of the maximum size of an earthquake for any given fault segment from geologic considerations alone. Such estimates are especially important in regions for which the maximum-magnitude earthquake may not be known from the available historical record. Moreover, the characteristic-earthquake model admits the interpretation that regional earthquake frequency-of-occurrence statistics are fundamentally controlled by the length distribution of coherent fault segments within that region, a proposition that has recently been verified for the island of Honshu, Japan. These developments indicate that much is still to be learned from the detailed mapping of active crustal fault zones to obtain coherent segment lengths and slip rates. This information, in turn, can be used to calculate much improved earthquake-risk estimates.

Equally plain are the manifestations that the earthquake rupture process itself is highly inhomogeneous. It has been known for some time that $\mathbf{M} \cong 8$ earthquakes, with faulting durations of 100 seconds or more, can be arbitrarily complex on 1-second time scales; so, too, $\mathbf{M} \cong 6\frac{1}{2}$ earthquakes can be arbitrarily complicated on 0.1-second time scales, and so on for even smaller earthquakes. It is natural to suspect that material inhomogeneities, as may be seen from geologic or seismicity evidence, play important roles in causing dynamically heterogeneous faulting. The rupture of high-strength asperities is now thought to play an important role in initiating crustal earthquakes—often dramatically—and barriers to faulting can produce large bursts of acceleration. Various types of earthquake series—for example, foreshock/mainshock/aftershock sequences or earthquake swarms—are now being modeled in terms of a distribution of inhomogeneous faulting elements.

It is difficult to overstate the importance of inhomogeneities in the Earth's crust with respect to the faulting process. Crustal earthquakes start and stop because of these material and geometric inhomogeneities, and what happens dynamically in between is also affected by them. All manner of geologic and seismologic phenomena are controlled by these inhomogeneities and their size/frequency-of-occurrence distribution. It is not too much to say that earthquakes would not occur as we presently know them, except for these inhomogeneous structures; indeed, but for them, earthquakes might not occur at all. Recent developments in the mapping and stochastic representation of these phenomena, both statically and dynamically, have led to much progress in understanding the causes and effects of earthquakes; it is certain that continuing research in these areas of multidisciplinary concerns will result in further advances.

## SPECIFIC ACCOMPLISHMENTS

### Crustal deformation

The Earth is a deformable body, and the ultimate cause of earthquakes is material deformation driven by relative displacements arising from large-scale plate motions. Beyond some critical strain level, this deformation can result in dynamically unstable shear failure, that is, an earthquake, although aseismic faulting and folding can and do occur as well. Geologic investigations of the long-term patterns and rates of crustal deformation have long held a central position in the earth sciences, for it is this deformation that shapes the face of the Earth in its great variety of geologic structures. Many of these structures, of course, are associated with economic, political, or some interest other than those important here, namely, geologic, tectonic, and seismologic structures associated with the causes and effects of crustal earthquakes.

In general, we need to know the sense of deformation (horizontal and vertical components, with their magnitudes), the patterns of deformation (both along strike and away from some fault of interest, for example), and the history of deformation. Recent developments in deciphering and dating geologic events in the late Quaternary (the past several hundred thousand years) have provided an important context for instrumental measurements. Studies of uplifted marine terraces in Washington, Oregon, and California, for example, have revealed that the Pacific coast of the coterminous 48 States has been uplifted substantially during the Quaternary. These uplift rates can exceed 10 mm/year in regions of crustal

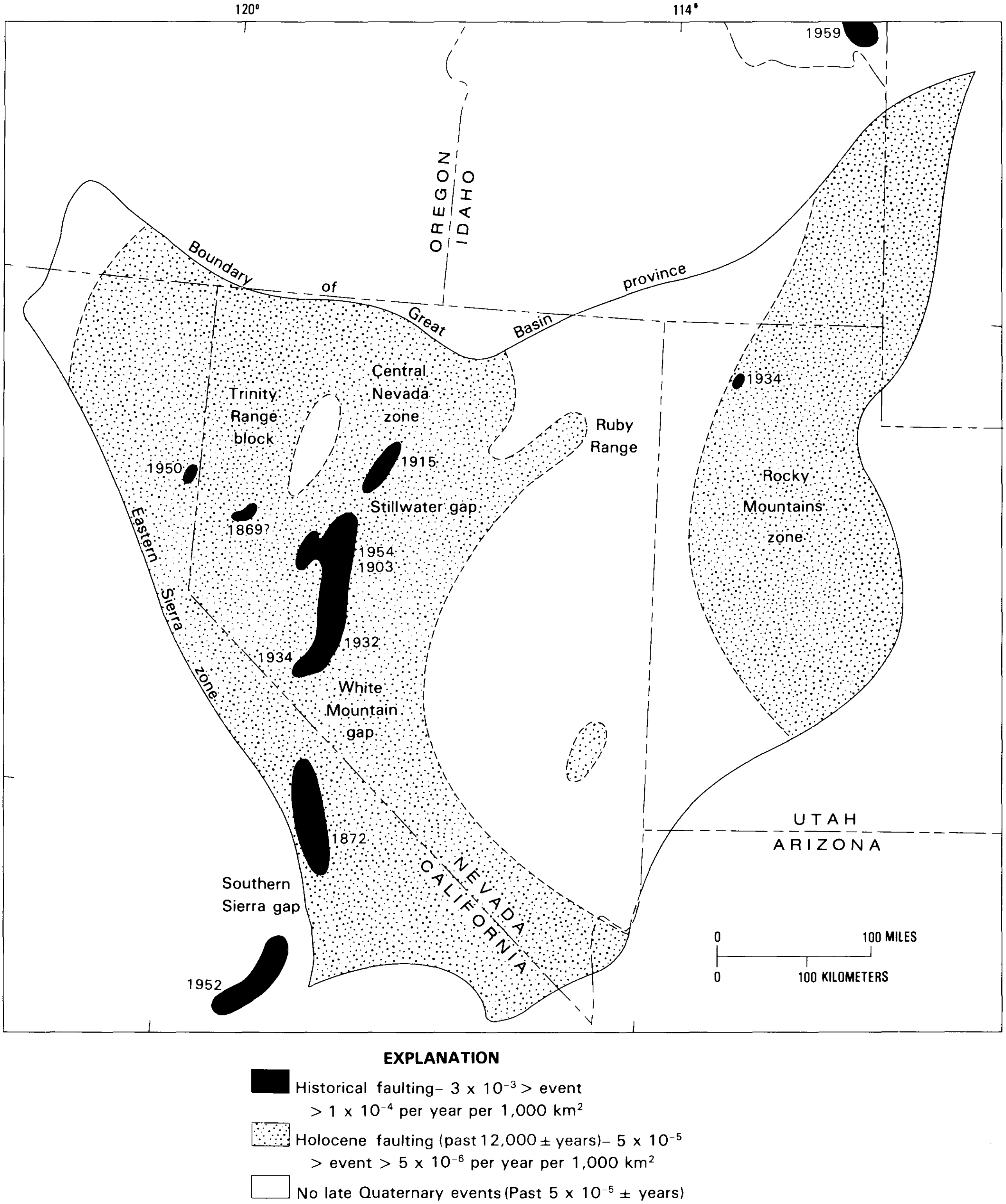
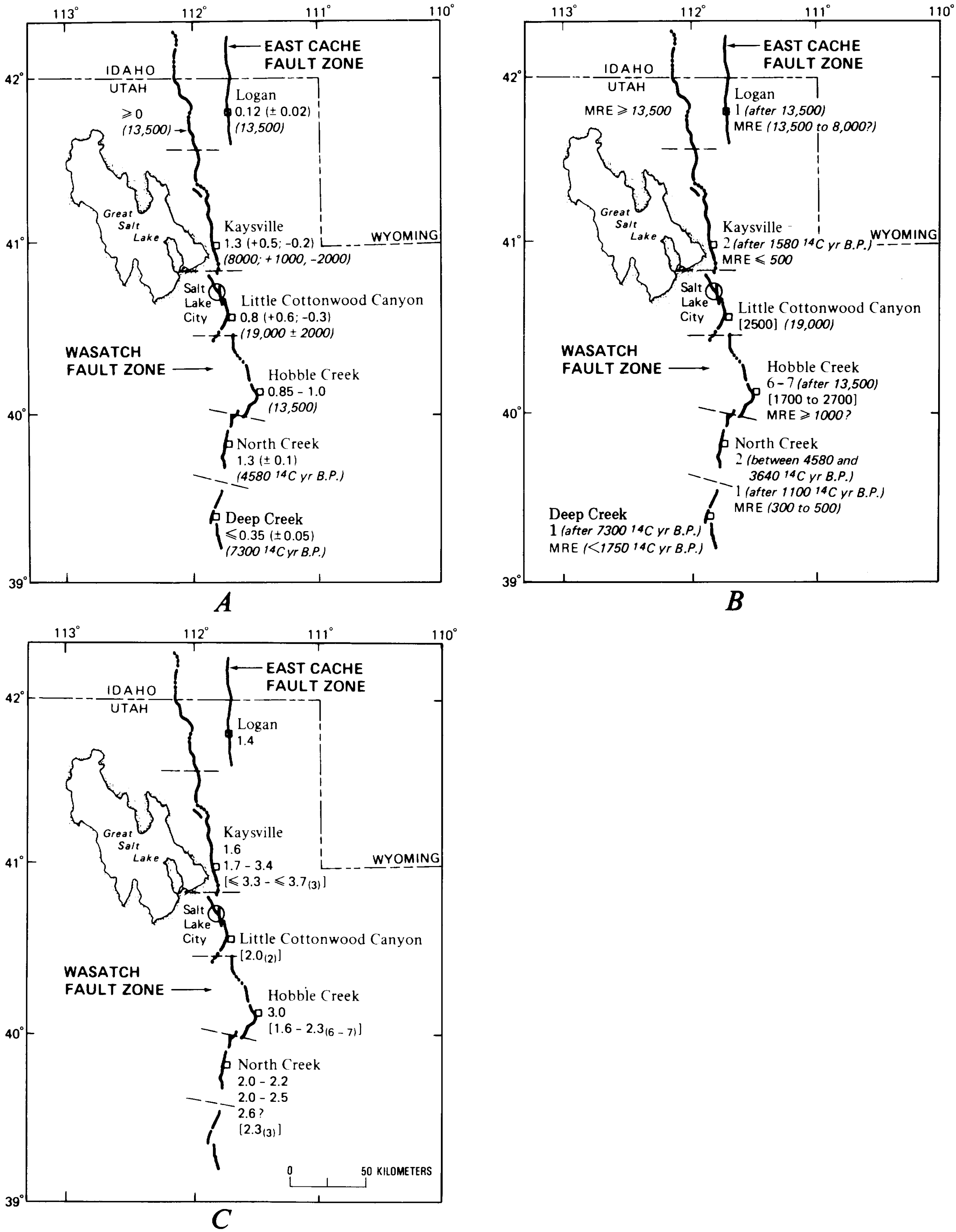



**Figure 8.** Generalized rates of surface faulting in the Great Basin province. From R.E. Wallace, "Active Faults, Paleoseismology, and Earthquake Hazards in the Western United States," *in* D.W. Simpson and P.G. Richards, eds., "Earthquake Prediction: An International Review (Washington, American Geophysical Union, p. 209–216), 1981.


113°
112°
111°
110°
42°
41°
40°
39°
EAST CACHE FAULT ZONE
IDAHO
UTAH
WYOMING
Great Salt Lake
Salt Lake City
WASATCH FAULT ZONE
≥0
(13,500)
Logan
0.12 (± 0.02)
(13,500)
Kaysville
1.3 (+0.5; -0.2)
(8000; + 1000, -2000)
Little Cottonwood Canyon
0.8 (+0.6; -0.3)
(19,000 ± 2000)
Hobble Creek
0.85 - 1.0
(13,500)
North Creek
1.3 (± 0.1)
(4580 14C yr B.P.)
Deep Creek
≤0.35 (±0.05)
(7300 14C yr B.P.)
A
MRE ≥ 13,500
Logan
1 (after 13,500)
MRE (13,500 to 8,000?)
Kaysville
2 (after 1580 14C yr B.P.)
MRE ≤ 500
Little Cottonwood Canyon
[2500] (19,000)
Hobble Creek
6 - 7 (after 13,500)
[1700 to 2700]
MRE ≥ 1000?
North Creek
2 (between 4580 and 3640 14C yr B.P.)
1 (after 1100 14C yr B.P.)
MRE (300 to 500)
Deep Creek
1 (after 7300 14C yr B.P.)
MRE (<1750 14C yr B.P.)
B
Logan
1.4
Kaysville
1.6
1.7 - 3.4
[≤ 3.3 - ≤ 3.7(3)]
Little Cottonwood Canyon
[2.0(2)]
Hobble Creek
3.0
[1.6 - 2.3(6 - 7)]
North Creek
2.0 - 2.2
2.0 - 2.5
2.6?
[2.3(3)]
0
50 KILOMETERS
C

shortening, such as the Ventura-Santa Barbara area on the southern margin of the Transverse Ranges, but even in regions dominated by strike-slip faulting, uplift rates are several tenths of a millimeter per year. In view of these observations, the Palmdale uplift in southern California is, geologically speaking, no surprise, although charting very short period (years), episodic behavior requires instrumental measurements (fig. 10). Further examples of the use of the recent geologic record to derive patterns and rates of deformation are given in the next section.

Regional geodetic measurements can determine modern rates of crustal deformation driven by longer term geologic processes. Figure 11 shows the principal strain rates at selected trilateration networks of the western United States. The modern seismic and tectonic activity of this region exists largely because the relative displacement between the North American and Pacific plates is not entirely accommodated at the principal plate boundaries along the Pacific coast. The observations in figure 11 yield estimates of this "missing" component of plate motion, thereby providing a quantitative basis for assessing the earthquake potential of this large intraplate region.

The most complete geodetic measurements in tectonically active areas come from California and Japan, where areal coverage is extensive, the history of measurements is long, and present-day measurements are made annually or more frequently. Along the San Andreas fault, for example, strain accumulation now occurs along the two locked segments, those which broke in the great earthquakes of 1857 and 1906; little if any strain accumulation is taking

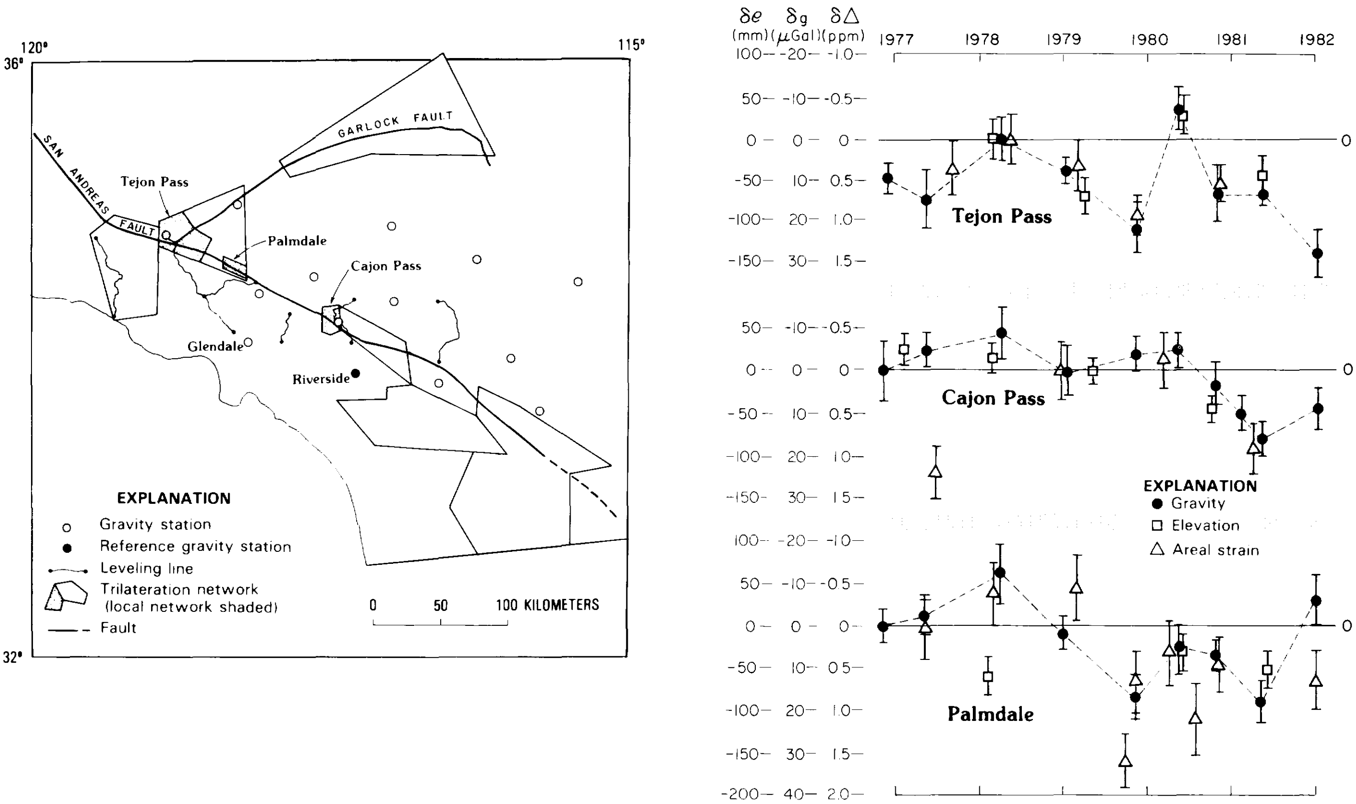



**Figure 10.** *Left,* index map showing locations of gravity stations, leveling baselines, and trilateration networks in southern California that have been surveyed repeatedly during the past 5 to 10 years. *Right,* temporal changes in gravity (δg), elevation (δe), and areal strain (δΔ) in three areas of southern California (shaded on index map). From R.C. Jachens, W. Thatcher, C.W. Roberts, and R.S. Stein, "Correlation of Changes in Gravity, Elevation, and Strain in Southern California" (Science, v. 219, p. 1215–1217), 1983.

**Figure 9.** Summary of slip rates, earthquake-recurrence times, and displacements per event for the Wasatch and East Cache fault zones in Utah and Idaho. **A,** Slip rate in millimeters per year; age of displaced datum (in italics) in years before present. **B,** Leading numeral gives number of surface-faulting events (after date, in years before present); bracketed numbers give average recurrence time (in years). **MRE,** most recent event, with elapsed age (in years) in parentheses. **C,** Displacement per event (in meters); brackets contain average displacement per event for number of events, in subscripted parentheses. From D.P. Schwartz, K.L. Hanson, and F.H. Swan III, "Paleoseismic Investigations along the Wasatch Fault Zone: An Update," *in* A.J. Crone, ed., "Paleoseismicity Along the Wasatch Front and Adjacent Areas, Central Utah" (Utah Geological and Mineral Survey Special Studies, v. 62, p. 45–49), 1983.

place along the intervening creeping section of the fault between Parkfield and San Juan Bautista. Data from high-precision surveys of the past decade can now be used to obtain accurate estimates of present-day slip rates along the major strands of the San Andreas fault, measurements that are now being complemented by long-term estimates extracted from the geologic record (see next section). Various satellite and astronomic distance-measuring techniques, developed in research programs conducted by the U.S. National Aeronautics and Space Administration, are now providing deformation data on the distance scales of the plates themselves.

Important recent achievements in understanding crustal deformation relate to the identification of significant departures from uniformity in the rate of strain accumulation. The existence, for example, of postseismic transients that are typically 20 to 40 percent of the coseismic strain drop reveals a nonconstant rate of strain accumulation; these results, in turn, have implications both for the mechanism of strain buildup and for the estimation of recurrence times. Evidence for these transients from the San Andreas fault is summarized in figure 12. These data demonstrate the large magnitude of the postseismic strains, but, because of the infrequency of postearthquake surveys, their duration is not constrained. Observations from great Japanese thrust earthquakes are more complete and indicate the existence of both short-term ($\cong$1 year) and long-term (decades) transients, results that point to the importance of both aseismic fault slip and transient flow in the asthenosphere for the cyclic accumulation and release of stress at major plate boundaries.

In addition to these systematic postseismic transients in strain accumulation, it is now known that strain accumulation can be accompanied by remarkable episodicity in deformation rate, as figure 10 clearly indicates for the rise and fall of southern California. The mechanisms causing these sporadic (but correlated) changes in gravity, elevation, and areal strain are not yet known. They do, however, indicate that tectonically driven deformation can bump along with disconcerting short-term irregularity, in the absence of significant earthquakes.

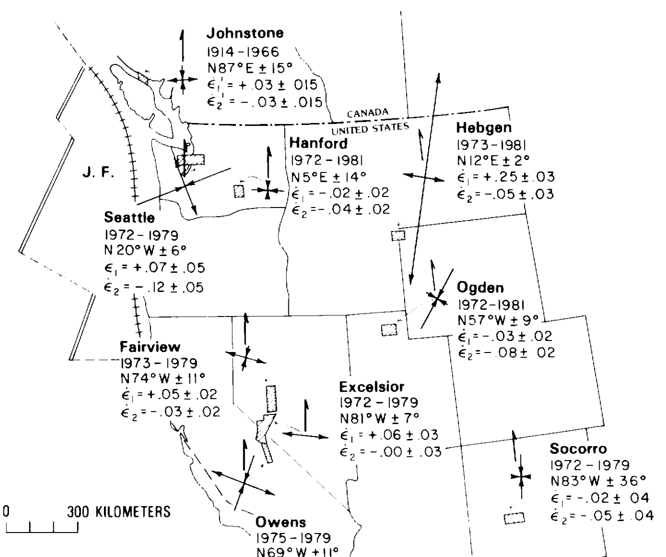



**Figure 11.** Trilateration networks (shaded areas) in the western United States and principal strain rates measured in each. Beneath each network name are the timespans of observations, directions of maximum extension, and principal strain rates (in microstrain per year). Major plate boundaries along the Pacific coast are sketched as dashed line (transform), double line (ridge), and shaded line (trench). **J.F.,** Juan de Fuca plate. From J.C. Savage, "Strain Accumulation in Western United States" (Annual Review of Earth and Planetary Sciences, no. 11, p. 11–43), 1983.

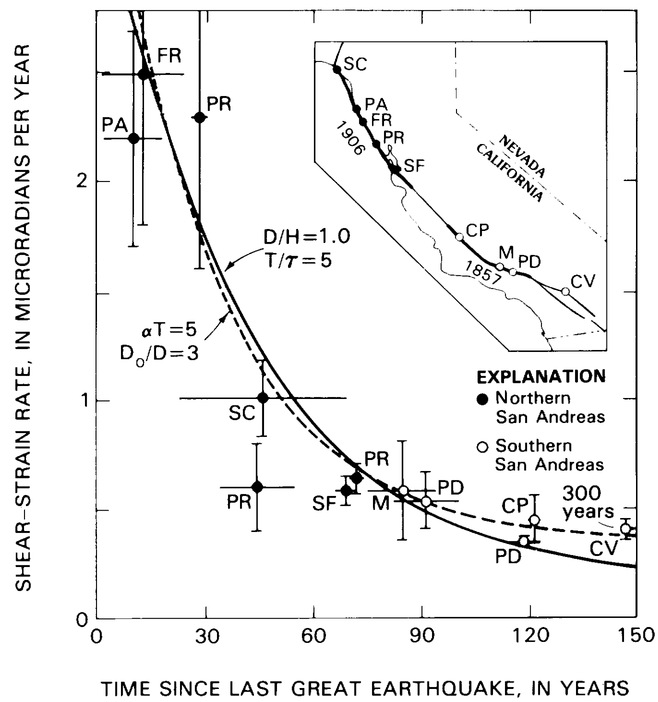



**Figure 12.** Shear-strain rate as a function of time since the last great earthquake on the San Andreas fault, either the 1857 Fort Tejon or 1906 San Francisco earthquake (see inset). Data from local networks: **CP,** Carrizo Plain; **CV,** Coachella Valley; **FR,** Fort Ross; **M,** Mojave Desert; **PA,** Point Arena; **PD,** Palmdale; **PR,** Point Reyes; **SC,** Shelter Cove; **SF,** San Francisco peninsula. Solid curve is a model calculation of temporal decay of strain rate for a coupled lithosphere-asthenosphere pulsed by earthquake loading; dashed curve shows calculation based on a modified elastic-half-space model. From W. Thatcher, "Nonlinear Strain Buildup and the Earthquake Cycle on the San Andreas Fault" (Journal of Geophysical Research, v. 88, p. 5893–5902), 1983.

## Late Quaternary faulting geology

The greatest difficulty imposed on earthquake studies is that of the available historical record of earthquakes. In California, for example, the record of felt earthquakes before the influx of miners and settlers beginning in 1849 can only be regarded as spotty; before the founding of Mission San Diego de Alcalá in 1769, it is nonexistent. Even in China, where the history of recorded earthquakes dates back some 3,000 years, statistical analysis suggests that the historical record is unlikely to be complete before A.D. 1500. As prisoners of the historical record, we are not only highly uncertain about the average recurrence intervals of large earthquakes along major plate boundaries but, in many less seismically prone areas, virtually ignorant of whether major earthquakes might occur every several hundred years—or every several hundred thousand.

An important development in earthquake studies, then, has been the extension of the short instrumental/historical record of earthquakes backward in time by as much as several tens of thousands of years. The techniques of "paleoseismology" are principally geologic in nature, although analytical dating techniques generally play a critical role in dating one or more key horizons or structures (below). The geologic investigations include analyses of faulting-controlled microstratigraphic relations, both along strike and in sections across the fault exposed by trenching; seismically induced sediment transport; fault-scarp and fault-trace geomorphology; the attitude of fluvial terraces progressively uplifted by faulting; and regional tectonic relations and syntheses.

Advances in dating late Quaternary materials have also progressed rapidly, and more than two dozen techniques are now being used. Uranium-series and $^{14}C$ methods have long been the mainstays of late Quaternary dating, when the appropriate material is available in sufficient quantity. A new technique of accelerator counting of $^{14}C$, however, allows one to work with very small samples of carbonaceous material. Much progress has been made in relative dating techniques, which will allow for absolute ages when properly calibrated against absolute-age scales. These include $^{10}Be$ accumulation in soils, the development of soil chronology and identification of paleosols, carbonate accumulation in soils, the development of hydration rinds, amino-acid racemization, dendrochronology, and lichenometry. Calibration of the extent of weathering against the bulk sound speed of exposed boulders shows much promise for determining the ages of the surfaces on which they lie.

The locality richest in evidence for repetitive earthquakes is Pallett Creek, California, where ponded sediment traversed by a segment of the San Andreas fault that ruptured in the 1857 earthquake has now been exposed by downcutting of the drainage since its capture in the early 1900's. Here, the faulting history of earthquakes dating back to the birth of Christ has been left in astonishing detail (fig. 13). This and other stratigraphic sections provide evidence of the occurrence of 12 earthquakes, comparable in size to the 1857 earthquake, during the past 1,700 years, with an average recurrence interval of approximately 150 years.

With observations that are extraordinarily easy to come by, the detailed morphology of dip-slip fault scarps is now being used to determine the time elapsed since their formation. A significant feature of this approach is that one is dating (or at least attempting to date) the earthquake itself, rather than some surface or stratigraphic horizon faulted at a later and generally unknown time. In essence, the proposition is very simple: for two scarps of comparable height in comparable material and climates, the older scarp is the one with the more subdued topography, in both slope and curvature. Recent applications of diffusion-equation mathematics allow quantitative determinations of such relative ages, and absolute ages in cases where the mass diffusivity can be estimated independently. Figure 14 illustrates how this technique has been used to analyze the faulting-controlled topography on which a section of the Great Wall in Ningxia-Hui Autonomous Region, People's Republic of China, was constructed. The wall, built here in A.D. 1531, was faulted in 1739; an earlier event with somewhat larger faulting displacements occurred along this same fault about 12,000 years ago.

Such investigations, both in the nature and scope of the geologic investigations and in improved, redundant age determinations, have contributed a great deal of information about average slip rates, number of events, and recurrence intervals between events along the fault segments being studied (fig. 9 represents another such study). Currently, some 150 geologic determinations of slip rates are available for California alone. These data have obvious and important implications for earthquake-risk analysis, as discussed in a later section, but these quantitative rates of deformation are also doing much to advance the scientific fields of Quaternary geology and tectonic geomorphology.

In thrust-faulting environments, near-surface sedimentary sequences on the upthrown block are commonly folded and arched upward, forming anticlines or monoclinal upwarps. Ancient faulting events may be recorded in warped fluvial terraces adjacent to streams that traverse such structures, as was the case for both the 1980 El Asnam, Algeria, and the 1983 Coalinga earthquakes. Recent events, of course, are recorded in reverse curvature of the stream profile itself. In the case of the El Asnam earthquake, faulting resulted in damming of the Chelif River and flooding of several square kilometers of the Middle Chelif Basin upstream from the upthrown block. Such "earthquake-induced reservoirs" are an unusual form of earthquake hazards, one that often can be put to good work.

Indeed, it is worth remembering that although faults, faulting, and earthquakes are recognized nowadays principally, if not exclusively, in terms of their hazard potential and nuisance value, it would be difficult to recognize many terranes, either geologically or economically, in the absence

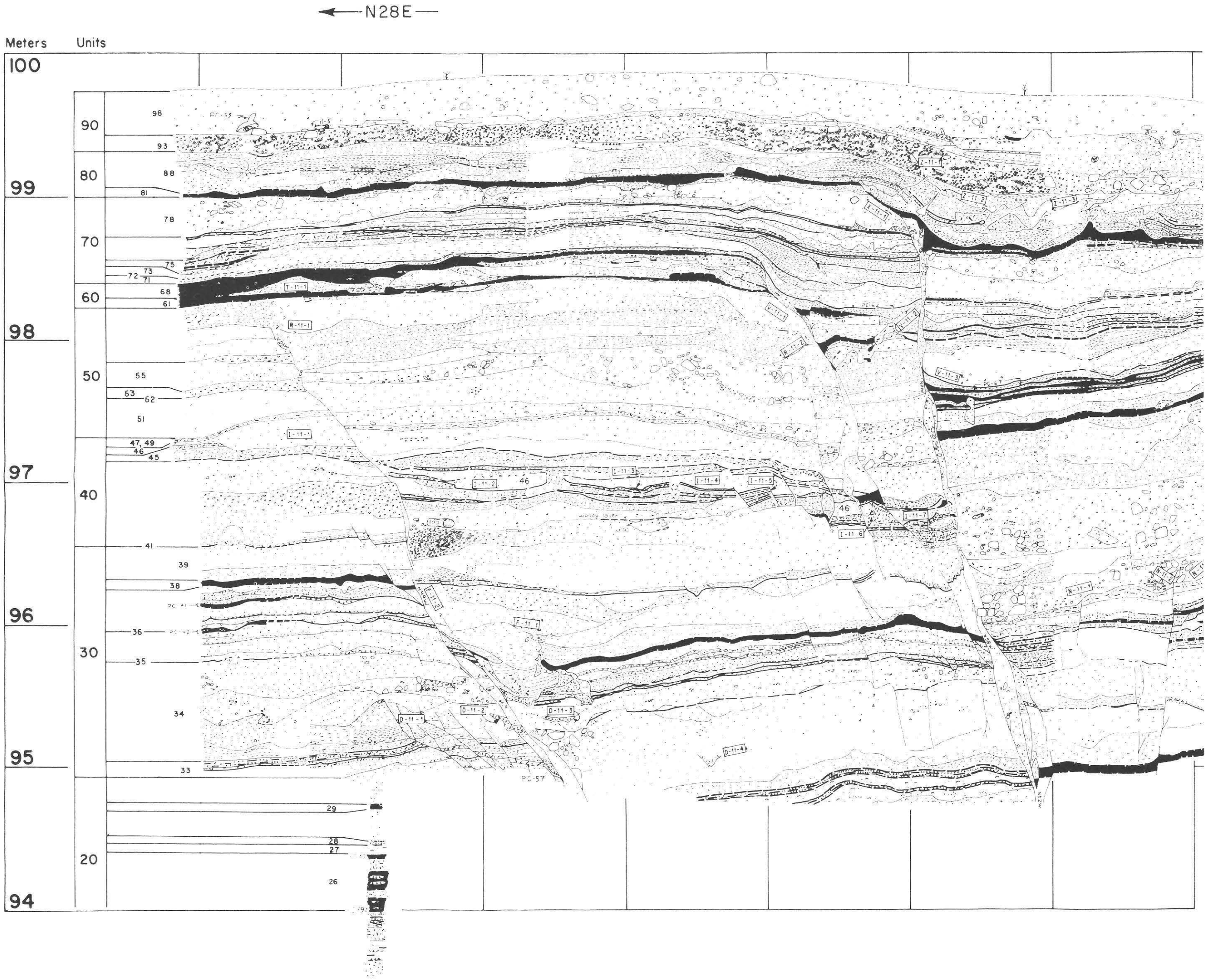



**Figure 13.** San Andreas fault exposed in a trench at Pallett Creek, California. Black horizons are datable peat layers, which are progressively offset by greater amounts at greater depths (greater ages), the cumulative effect of earthquake faulting. View southeastward. From K.E. Sieh, "Prehistoric Large Earthquakes Produced by Slip on the San Andreas Fault at Pallett Creek, California" (Journal of Geophysical Research, v. 83, p. 3907–3939), 1978.

of their effects. In California, for example, the Sierra Nevada as we presently see it, to which we owe the placer gold and much of the State's water supply, exists because of uplift and backtilting caused by spectacular normal faulting along its east front. The enormous sand-and-gravel operations in the Los Angeles Basin would operate at a fraction of their present production, if at all, but for the erosional debris delivered annually from the faulting-controlled San Gabriel Mountains standing just to the north. The Crystal Springs Reservoir, which holds Hetch-Hetchy water from the Sierra Nevada for use in San Francisco, owes its existence principally to the rift zone of the San Andreas fault. And the first Old World child born in California opened his eyes to the Coyote Creek fault on Christmas Eve of 1775, as the second Anza expedition to settle California availed itself of this strand of the San Jacinto fault zone on their journey from Yuma to the missions of the Los Angeles Basin.

## Seismicity studies

A fundamental aspect of earthquake studies is simply knowing where earthquakes occur (and where they will occur in the future), and so systematic location of earthquakes has been a central task of seismology since continuously recording seismographs were developed a hundred years ago. Another point of view is that seismicity studies, in their earthquake locations and focal mechanisms, provide us with an "instantaneous" picture of the release of crustal strain. In this section, we emphasize developments in situations for which there is a close association between seismicity and crustal fault zones in active tectonic environments. Equally

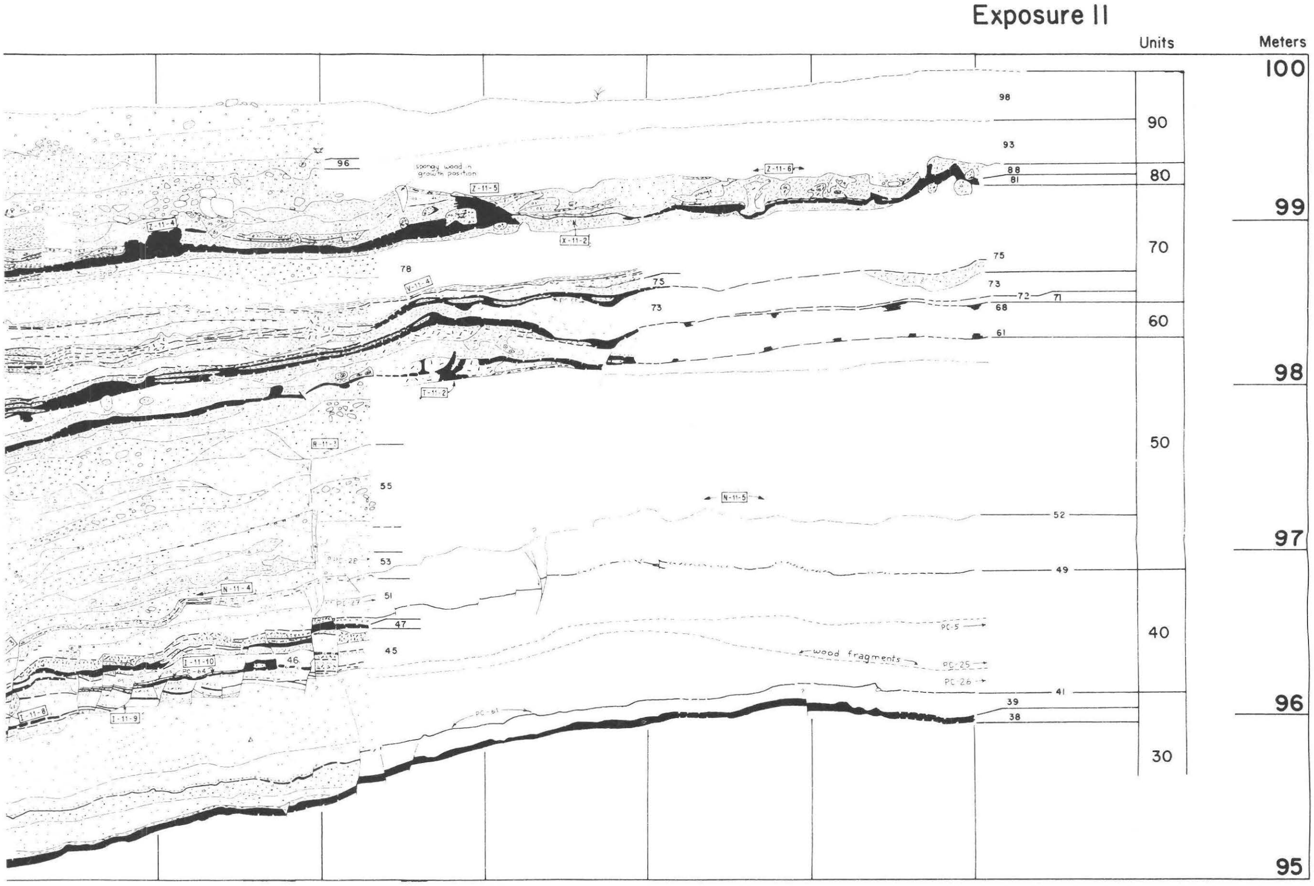



**Figure 13.** Continued.

important associations are now known between earthquakes and both volcanic processes and human activities; magmatic seismicity and induced seismicity are the subjects of the following two sections. After a one-section interlude on stress estimates, the important problem of seismicity in the eastern and central United States is taken up, for which the association between seismicity and geologic structure with surface expression is far less clear.

An important development of the 1960's was the documentation of the close association between earthquake locations and crustal fault zones that could be judged active on purely geologic grounds. Since that time, computer-aided data-processing techniques have dramatically increased both the completeness and accuracy of earthquake catalogues. Figure 15, for example, shows the seismicity of western Nevada, California, northern Baja California, and the adjacent offshore areas; depicted here are some 7,000 $\mathbf{M} \geqslant 1.5$ earthquakes occurring in 1980. Figure 15 reproduces the "close" association between seismicity and active crustal fault zones (compare with the block boundaries of fig. 7) but also reveals that this association is not exact. The sections of the San Andreas fault that broke in the 1857 and 1906 earthquakes, for example, are conspicuously aseismic, as is the section of fault that broke in the great 1872 earthquake along the east front of the Sierra Nevada. We now recognize all three of these fault segments as seismic gaps, in the quiescent stage of their earthquake cycles. On the other hand, the boundary between the Coast Ranges and the Great Valley east of the San Andreas fault is noticeably seismic, in the absence of surficial expression of throughgoing faults. The 1983 Coalinga earthquake occurred along this zone. Significantly, no surface faulting has been directly associated with this earthquake (although minor surface offsets developed during the aftershock sequence)—a rare situation for California earthquakes of this magnitude.

A perplexing question in earthquake studies and one of great significance for earthquake prediction involves the processes by which one earthquake may (or may not) condition the occurrence of yet another, possibly larger event, either in its location in space and time or, possibly, in its source parameters. It has been known for a long time, for example, that most $\mathbf{M} \gtrsim 6$ crustal earthquakes have aftershock sequences easily identifiable as such in terms of their spatial and temporal distributions. Detailed studies in recent years, especially those of aftershocks of the 1975 Oroville and 1979 Coyote Lake earthquakes, have further revealed that aftershocks can migrate well away from the surface faulted at the time of the main shock. These results are understood in terms of the elastic-stress increases on the fault exterior to the region of the main-shock rupture, itself now carrying a reduced shear stress. Progressive seismic failure then ensues as the more highly stressed exterior regions subsequently fail, at least locally, in the form of aftershocks. That is, the main shock drives the aftershock sequence with its own stress redistribution.

It has also been known for some time that most crustal earthquakes do not have such easily identifiable foreshock sequences; if such a "sequence" is identified at all, it need involve no more than one or two events. Almost certainly, the accumulation of stress leading to some main shock occurs much more slowly than the release and redistribution of stress attendant to the main shock, and it seems likely that "foreshock sequences" can only be defined over correspondingly longer periods of time. What does *not* seem likely is that these processes of strain accumulation would have no seismic expression whatsoever. In fact, several recent studies have indicated that localized zones of impending rupture can be outlined (in shadow) by the preceding long-term seismicity of the surrounding regions, and the 1983 Coalinga earthquake is one example. Nevertheless, much work remains to be done on what, if any, systematics (both in seismicity patterns and in source parameters) are associated with the seismogenic

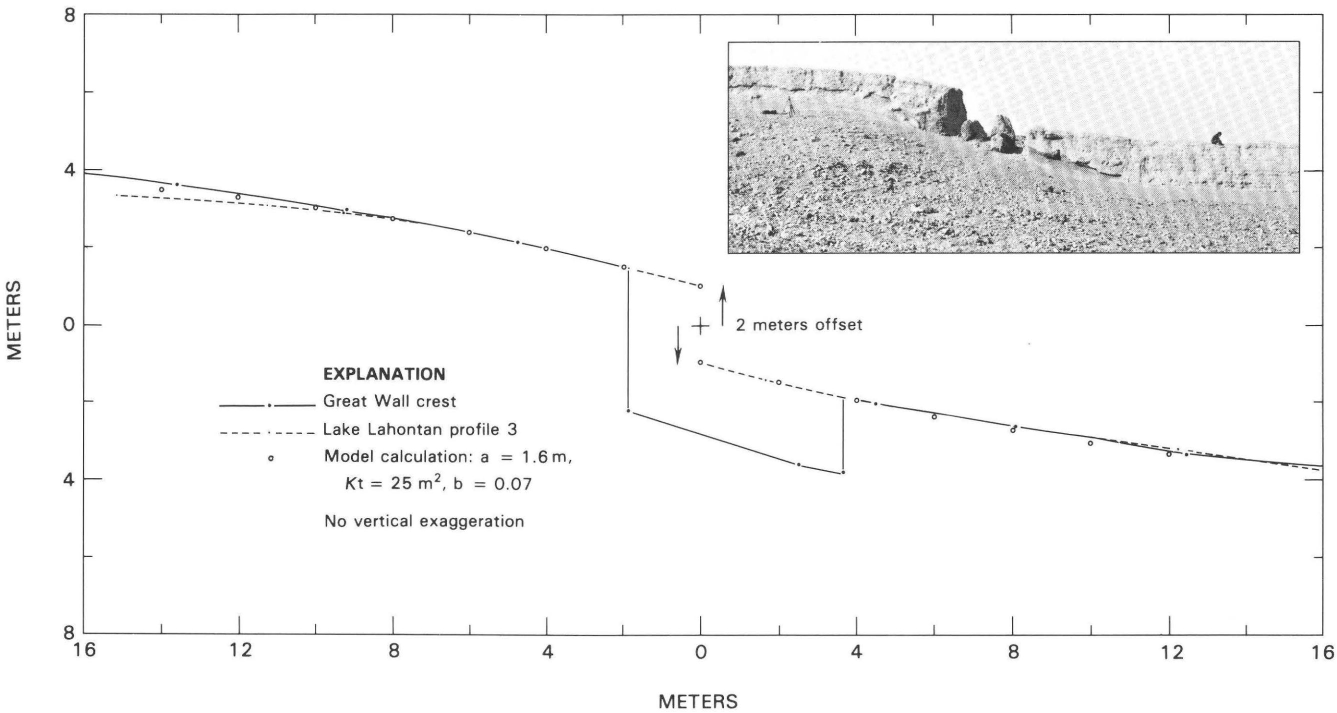



**Figure 14.** *Inset,* the Great Wall southwest of Shizuishan, Ningxia-Hui Autonomous Region, People's Republic of China. The 1739 earthquake faulting cuts obliquely across photograph, from damaged part of the wall to near center of left margin. Photograph by R.E. Wallace. *Below,* profile of crest of the Great Wall shown above, in comparison with a fictitiously offset profile of a Lake Lahontan shoreline scarp in Pershing County, Nevada. Shoreline scarp is approximately 12,000 years old. This comparison, together with "faulted" diffusion-equation calculations (dots), indicate that the Great Wall here was built on faulting-controlled topography approximately 12,000 years old. Faulting displacement of this earlier earthquake is about 60 percent larger than that for the 1739 earthquake at this locality. From R.C. Bucknam, T.C. Hanks, and R. E. Wallace (unpub. data, 1984).

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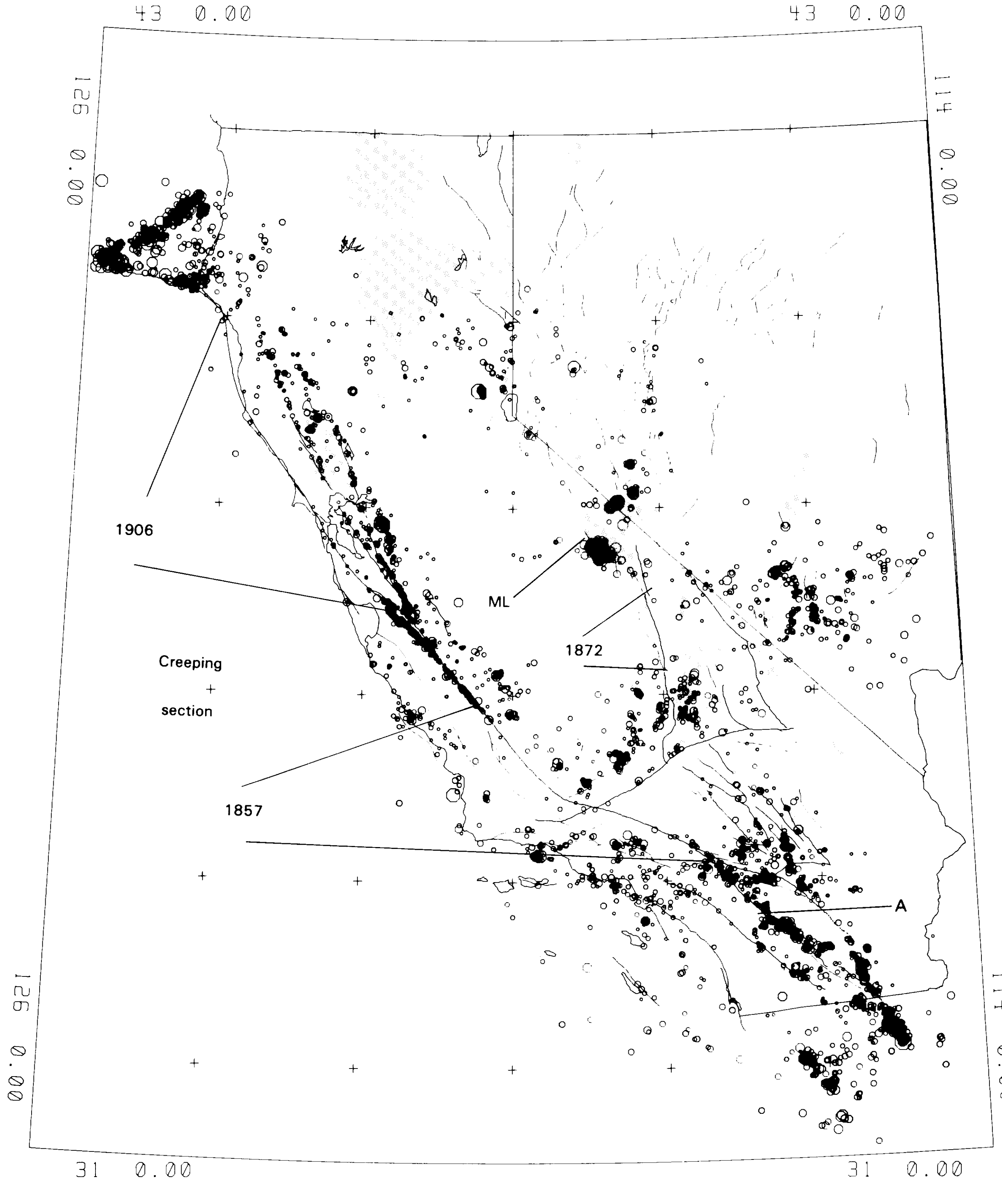
**Figure 15.** Seismicity of western Nevada and California in 1980 (**M** > 1.5). Smallest symbols, **M** < 2; largest symbols, **M** ≥ 7. Heavy lines, faults with Holocene activity; shaded areas, Quaternary volcanic centers. Straight-line end points denote extents of the 1906 San Francisco earthquake, rupturing the San Andreas fault from Shelter Cove to San Juan Bautista; creeping section of the San Andreas fault from San Juan Bautista to Parkfield-Cholame; the 1857 Fort Tejon earthquake, rupturing the San Andreas fault from Parkfield-Cholame to Cajon Pass; and the 1872 Owens Valley earthquake, breaking the Sierra front from Big Pine to the south end of Owens Lake. Mammoth Lakes and Anza are at end points of straight lines **ML** and **A,** respectively. Aftershocks of the 1980 Eureka earthquake define northeast-southwest zone of seismicity in upper left corner. Modified from D.P. Hill, R.E. Wallace, and R.S. Cockerham, "Review of Evidence on the Potential for Major Earthquakes and Volcanism in the Long Valley-Mono Craters-White Mountains Regions of Eastern California" (Earthquake Prediction Research, v. 3, p. 551–574), 1985.

expression of strain accumulation in regions specifically identified as likely candidates for **M** $\gtrsim$ 6 crustal earthquakes. In California, such experiments are in place near Parkfield and Anza.

## Magmatic seismicity

Nowhere is the association between seismic activity, crustal deformation, and natural process so dramatic as in the case of the emplacement of partially molten rocks into the



Earth's crust. As has been documented many times at Kilauea Volcano, Hawaii, magmatic emplacement at shallow depth leads to uplift and extension of the overlying crustal rocks, deformation easily monitored by leveling and geodetic techniques. More rapid and localized motion of magma, especially as it approaches the Earth's surface, results in earthquake swarms (with rates of activity commonly exceeding 1,000 events per day) and harmonic tremor. Many of these processes, as revealed by their characteristic seismic and geodetic signatures, culminate in volcanic eruptions, numerously at Kilauea and spectacularly at Mount St. Helens, Washington, but many others do not, or have not yet, as in the cases of Matsushiro, Japan, and Long Valley, California.

Although such investigations are not written specifically into the mandate of NEHRP, the early recognition of Mount St. Helens' awakening from dormancy and the systematic prediction of eruptions subsequent to the May 18, 1980, explosion, based chiefly on the rapid analysis of data from a local seismograph network, are signal seismologic accomplishments. The past two decades have witnessed close associations between students of volcanic and tectonic earthquakes. The high-gain, short-period telemetered seismograph networks used extensively in California, Nevada, Utah, and Washington—indeed, in many parts of the world—trace their heritage to instruments developed during the 1950's to study microearthquakes on Kilauea; conversely, many of the seismologic capabilities that served so well at Mount St. Helens were developed in response to NEHRP. In any case, real or imagined programmatic distinctions lose their meaning in such areas as the Imperial Valley, Coso Hot Springs, and Long Valley, where tectonic and magmatic processes are intimately related. Around the Pacific margin, volcanic studies have also been used to develop stress trajectories, and there is occasional evidence that occurrences of major volcanic eruptions along the "Rim of Fire" are related to occurrences of adjacent great earthquakes.

Perhaps more important, however, are the close parallels between earthquake prediction and the prediction of volcanic eruptions, both as matters of scientific method and as matters eliciting significant economic, political, and social responses. With respect to the scientific parallels, the principal techniques in both cases are the same: measurements of seismic activity and crustal deformation, and observations of other phenomena, such as gas emanation, that relate indirectly to either or both. Anomalous seismic and geodetic activity that precedes volcanic activity is of much interest to earthquake-prediction studies, both because such anomalous premonitory volcanic activity is commonly associated with very large signals and because many volcanoes, such as Kilauea, have very short interevent times.

Responding to some level of notice, warning, alert, or prediction, however, is another matter, especially in cases when the event of interest (an earthquake or an eruption—or possibly both) may be long delayed from the onset of premonitory activity. The situation at Long Valley that has developed since 1978 is a case in point. Southern Mono County, with Mammoth Lakes in its southwest corner, has been the locus of volcanic activity for millions of years. Long Valley caldera is the result of an enormous volcanic explosion 700,000 years ago, and smaller surface eruptions have continued almost to the present, the latest being some 500 years ago. Against a backdrop of punctuated but more or less continuous earthquake-swarm activity since 1978 and 40 cm of uplift across Long Valley caldera in the period 1979 to the present, the USGS issued a Volcanic Hazards Notice for the area in May 1982. The intense swarm of January 1983 along the southwest caldera boundary, together with increasing fumarole activity, did nothing to diminish concern. Scientifically, there is agreement that both the seismic activity and the vertical uplift indicate mobilized magma at depths of 4 to 8 km (fig. 16). None of the scientists concerned would be surprised if this activity were to accelerate and result in a hazardous eruption next week, but neither would they be surprised if it were not to occur for another hundred years. While the impending hazard is real, then, its timing is highly uncertain. As chronicled by George Alexander of the Los Angeles Times, such a situation presents unusual difficulties in constructing a measured and prudent response that is both scientifically and politically viable.

## Induced seismicity

The spatial and temporal occurrence of earthquakes has been influenced by human works in a number of ways, including the filling of large reservoirs, injection of pressurized fluids into deep boreholes for waste-disposal purposes, fluid injection for secondary recovery of hydrocarbons at depth, fluid transport related to solution mining and geothermal-energy extraction, and both underground and surface mining. The largest induced earthquakes have occurred in close spatial association with large reservoirs; earthquakes of $\mathbf{M} \geqslant 5\frac{1}{2}$ have occurred at Koyna, India; Kariba, Zambia/Zimbabwe; Kremasta, Greece; Oroville, California; Hsinfengkiang, China; and Aswan, Egypt. The largest earthquake to have been associated with mining operations is the $\mathbf{M} = 5.5$ earthquake in the Klerksdorp Mining District, South Africa, although concern exists there that even larger events may occur as gold mining progresses beyond the 4,000-meter depth level.

Although the association between reservoirs and enhanced seismic activity had been noted as long ago as the filling of Lake Mead behind Hoover Dam, the physical mechanism of this process and its potential for generating damaging and destructive earthquakes have been appreciated only recently, specifically through analyses of earthquakes induced by deep-level waste disposal (1962–64) at the Rocky Mountain Arsenal near Denver, Colorado, and by a controlled fluid-injection experiment (1970–74) at the Rangely oil field in western Colorado. It is now understood that these earthquakes were the result of effective stress decreases caused by fluid-pressure increases, reducing the *in situ* frictional resistance to shear failure. Similarly, mining-induced earthquakes appear to be directly related to shear-stress buildup just ahead of advancing mine faces. These results have important implications for the relation between tectonic stresses and crustal deformation in general and the frictional (stick slip) model of crustal earthquakes in particular, subjects we take up in the next section.

Since 1975, field studies of reservoir-induced seismicity in Soviet Central Asia (Nurek and Toktogul Reservoirs) and Egypt (Aswan) have provided detailed case histories in different geologic and tectonic settings. These studies have shown that the potential for induced earthquakes depends strongly on localized geologic and hydrologic conditions. At the Nurek Reservoir, for example, induced earthquakes are concentrated in an area where a plunging anticline allows hydraulic connection between the reservoir and hypocentral depths. Few earthquakes occur in an immediately adjacent region, beneath an even greater reservoir depth; there, a tightly folded, horizontal syncline confines the influence of the reservoir to shallow depths.

The situation at the Monticello Reservoir in South Carolina, constructed to provide cooling water for the nearby Virgil C. Summer Nuclear Power Station, presented several unusual characteristics with respect to "normative" conditions of reservoir-induced seismicity, although the association between reservoir filling and earthquake rate of occurrence is clear (fig. 17). In the first place, the reservoir itself is small and shallow; secondly, the induced seismicity is confined to very shallow depth ($\lesssim 1.5$ km). Such depths are easily reached by drilling, and two boreholes penetrated the seismogenic zones to depths of $\cong 1.1$ km. In each hole, several suites of geophysical measurements were made, including *in situ* stress magnitudes, pore pressure, permeability, and the distribution and orientations of faults, fractures, and dikes.

Several significant conclusions may be inferred from these results and those of the companion seismicity and source-mechanism studies. First, the hydraulic diffusivity estimates permit the size of the seismically active zone to be

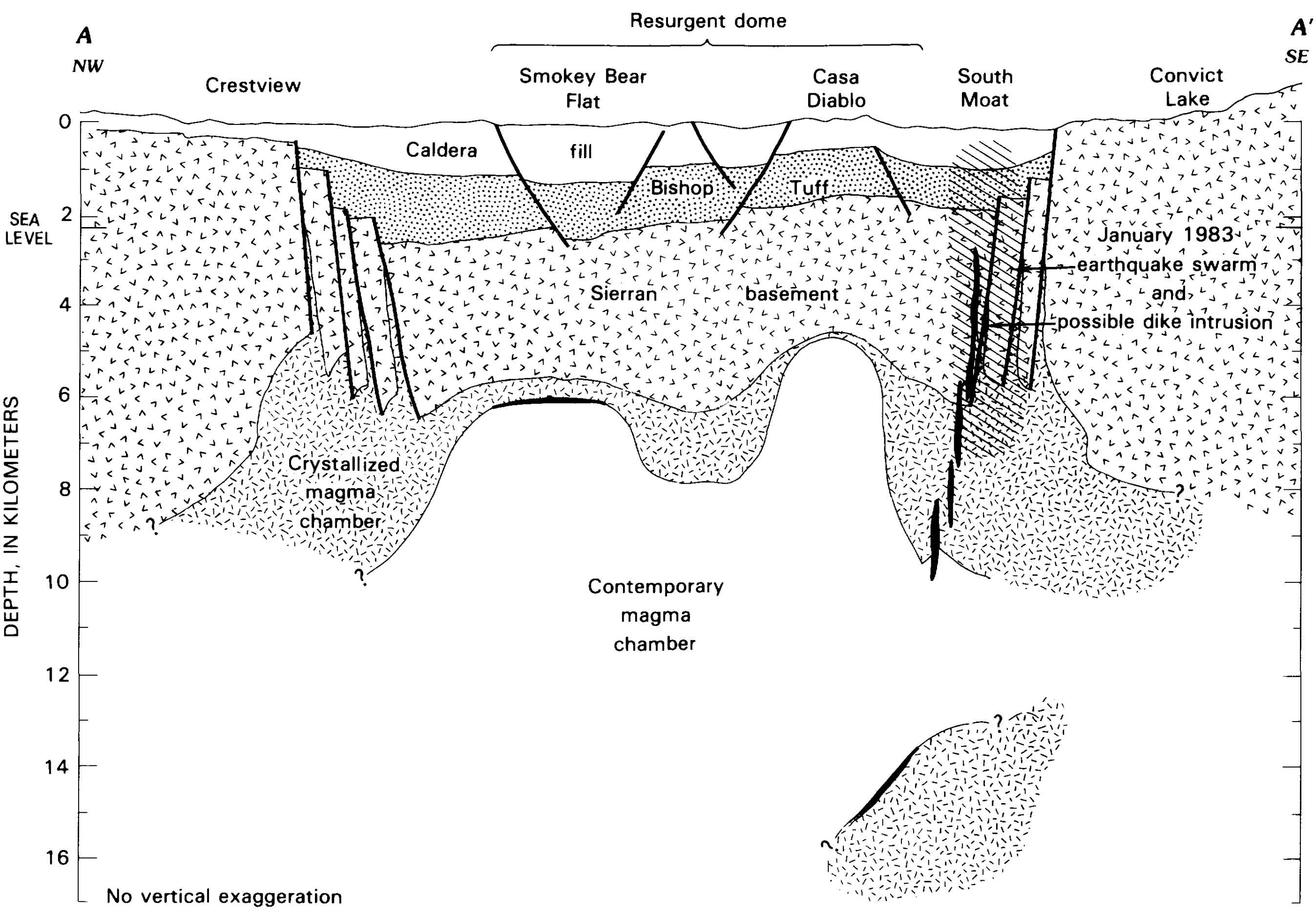



**Figure 16.** Northwest-southeast cross section through Long Valley caldera, showing principal geologic units. The town of Mammoth Lakes, California, is located near **A'**. Geodetic and seismic data suggest magmatic intrusion beneath south moat of the caldera. From D.P. Hill, R.E. Wallace, and R.S. Cockerham, "Review of Evidence on the Potential for Major Earthquakes and Volcanism in the Long Valley-Mono Craters-White Mountains Regions of Eastern California" (Earthquake Prediction Research, v. 3, p. 551–574), 1985.

the result of fluid diffusion alone. Second, the stress magnitudes reveal that the Monticello Reservoir area is only marginally stable to stick-slip (seismogenic) frictional failure. Third, the stress magnitudes here are nevertheless concordant with those obtained in other areas, otherwise considered to be tectonically stable and seismically inactive. All three of these conclusions point to a fourth: the seismicity at the Monticello Reservoir was, indeed, driven by the perturbing action of the reservoir itself.

## Stress measurements

The forces that drive plate motions, faulting displacements, and earthquakes result not only in deformation of the Earth's crust but also in deviatoric (shear) stresses within it. From the point of view of earthquake-mechanism studies, matters turn on whether or not crustal earthquakes result from stress-conditioned failure processes and, if so, what magnitude of deviatoric stress is necessary to drive that failure process. The time-predictable model of earthquake recurrence, for example, explicitly supposes a time-invariant level of critical shear stress for failure. Over the past two decades, laboratory investigations of the failure mechanics of precut rock masses have suggested that crustal earthquakes are controlled by frictional processes and their related stress systems. An important result of these experiments, formulated in the late 1970's as Byerlee's law, is that crustal rocks have nearly identical coefficients of friction, 0.85 at low confining pressures ($\sigma_n \lesssim 0.2$ GPa) and 0.6 at greater $\sigma_n$. (At depths greater than $\cong 15$ km in tectonically active areas other than subduction zones, deformation passes from the brittle-failure regime into temperature-dependent, ductile and plastic regimes; the deformation at these depths is largely, if not wholly, aseismic.)

Verifying the applicability of the laboratory friction results to the real Earth is a different and difficult problem. If we knew the crustal deformation field and its history in some detail and if we knew the appropriate rheologic laws for the long-term stress-strain behavior of crustal rocks, we could simply calculate the stress field. Neither supposition holds sufficiently well for this to be a fruitful exercise. The results from the Rocky Mountain Arsenal and Rangely, however, clearly suggest the applicability of the frictional model of faulting, at least to shallow depths ($\cong 5$ km); these experiments have not, unfortunately, resolved the problem of shear-stress magnitudes. That is, the driving tectonic stresses can be arbitrarily small if the local fluid pressure is arbitrarily close to the lithostatic pressure of the overlying rock. Geologic investigations of remanent frictional heating, expressed in anomalous, geometrically controlled metamorphic gradients and patterns of argon retention, are promising ways to infer shear-stress magnitudes, but such investigations are not likely to provide a large areal sampling. On the other hand, heat-flow measurements in California argue against significant shear heating (frictional stresses $\gtrsim 10$ MPa) along the San Andreas fault or any of its principal branches.

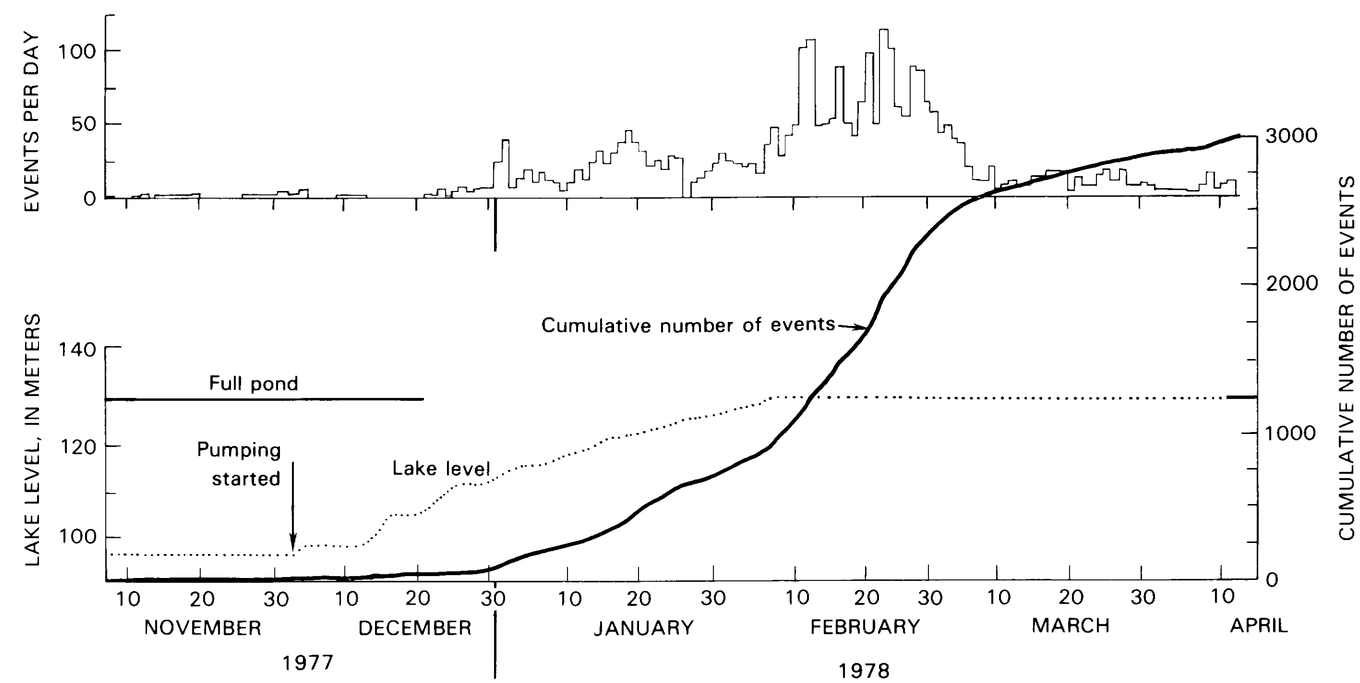



**Figure 17.** Number of earthquakes per day, cumulative number of earthquakes, and water level in the Monticello Reservoir, South Carolina, as functions of time. Note near-coincidence of onset times of reservoir filling and enhanced rate of seismic activity. From J.B. Fletcher, "A Comparison Between the Tectonic Stress Measured in Situ and Stress Parameters from Induced Seismicity at Monticello Reservoir, South Carolina" (Journal of Geophysical Research, v. 87, p. 6931–6944), 1982.

In these circumstances, the application of hydrofracture techniques to determine, directly and *in situ*, the principal stress magnitudes and orientations as a function of depth in tectonically active areas has been a welcome development. Since the inception of NEHRP, stress measurements have been made at Monticello, South Carolina, as discussed in the previous section, to illuminate the nature of shallow reservoir-induced seismicity at this locale; at Auburn, New York, to determine the proximity to shear failure and the earthquake potential in this largely aseismic upstate area; and along the Wasatch front, Utah, and at the Nevada Test Site to measure stress at depth in these Basin and Range settings. The Nevada Test Site data suggest that, at least to 1.3-km depth, this region is metastable with respect to normal faulting, if the failure process is that of the frictional model.

The most complete and telling data in this regard come from the Mojave Desert, adjacent to a section of the San Andreas fault which ruptured in the 1857 earthquake. The 29 measurements summarized in figure 18 were taken in a series of drill holes at distances between 2 and 34 km from the fault, to depths of 0.85 km. The depth coefficient of 7.86 MPa/km of the fitted straight line corresponds to a coefficient of friction of 0.45, with the assumption of a hydrostatic fluid pressure. This value is somewhat low in comparison with that calculated from Byerlee's law, but it is nevertheless consistent with frictional coefficients obtained for gouge taken directly from the San Andreas fault zone.

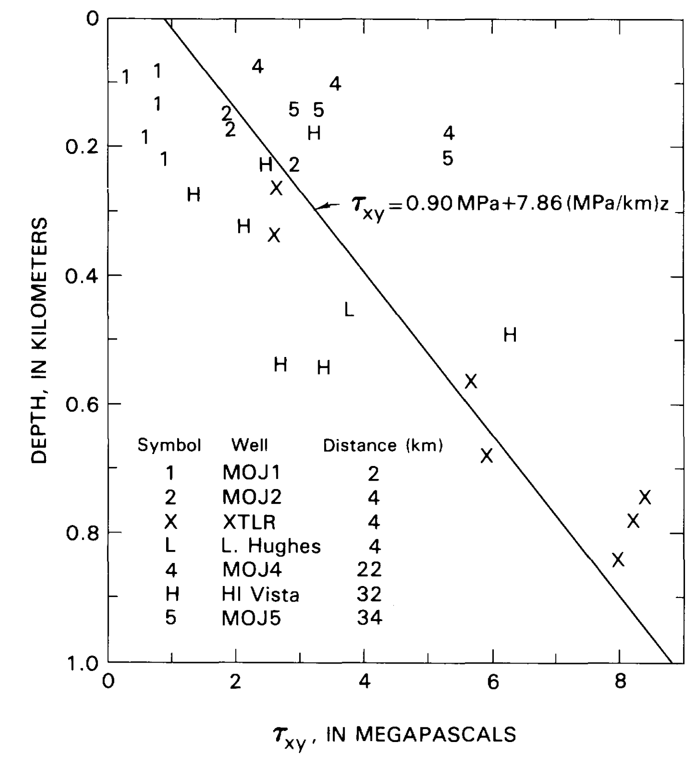



**Figure 18.** Horizontal shear stress as a function of depth near Palmdale, California; distances are normal to the San Andreas fault. From A. McGarr, M.D. Zoback, and T.C. Hanks, "Implications of an Elastic Analysis of in Situ Stress Measurements near the San Andreas Fault" (Journal of Geophysical Research, v. 87, p. 7797–7806), 1982.

Extrapolation of the results in figure 18 through the seismogenic depth ($\cong$ 15 km) yields shear stresses at the 0.1-GPa level. Even the average stress of 56 MPa for the upper 14 km exceeds the shear stresses permitted by the heat-flow data, by a factor of 3 or more. Either some mechanism acts at depth to distribute heat far more efficiently than thermal conduction alone (presumably, water circulation), or shear stresses on the San Andreas fault must level off quickly at depths below 1 km. Stress measurements as a whole set, however, are not known for such behavior; observations at even the deepest levels, in excess of 3 km, fail to show any suggestion of decrease in depth gradient.

Even in the absence of knowledge of stress magnitudes, the orientations of the principal stress axes contain much useful information on the contemporary forces acting on the Earth's crust. Figure 19 shows a recent compilation of such data for North America. The observations consist of alignments of geologically young ($\leq$ 2 million years) volcanic dikes and cinder cones, fault-slip observations, earthquake focal mechanisms, and wellbore breakouts. The general coincidence between coherent stress domains and well-defined heat-flow provinces suggests an underlying dynamic process (convection) beneath the brittle crustal layer. On a larger scale, the relatively uniform, northeast-directed compression observed in the central and southern United States contrasts markedly with the complex pattern of stress directions in the western United States, almost certainly the result of the Pacific-North American plate interactions in the latter region. Figure 19, then, is also the embryo of a tectonic regionalization of the United States, containing especially valuable information for earthquake hazards and risk assessments in the relatively aseismic eastern and central United States.

## Earthquakes in the "eastern" United States

Evaluating seismic hazards and earthquake potential in the United States east of the Rocky Mountains is complicated by the relative infrequency of potentially damaging and destructive earthquakes and by the general absence of the associations between seismicity and geologic structures with surface expression that are so clear in the western United States. It is also known that ground-motion attenuation with distance is far less in the East than in the West, and there is yet concern that earthquake source parameters, at the same $M_0$ or **M**, may differ in the two regions as well. The hazards and risks from earthquakes in the eastern United States are nonetheless real for these difficulties: damaging and destructive earthquakes have occurred widely through this large region (fig. 2); population densities are high, especially along the eastern seaboard; many of the major population centers have heavy concentrations of structures that are ancient and dangerous by modern standards; and most of the Nation's nuclear reactors are sited here as well.

The cause—or causes—of the diffuse, low-level seismicity in the eastern regions is not understood. As mentioned above in the subsection entitled "Structural and Tectonic Setting," an important recent development in the understanding of seismicity of the eastern United States has been the accumulating evidence that much of it is associated with ancient zones of weakness, reactivated by modern stress regimes. It is now thought, for example, that earthquakes along the Atlantic Coastal Plain are associated with boundary grabens formed upon opening of the Atlantic during Triassic time. The problem is that little is known about the underlying causes of the "modern stress regimes." It can be said that the seismic deformation rates are very low and that, by consequence, elastic-strain energy need be delivered to the eastern region at correspondingly low rates, in comparison with the western region say, but this observation almost certainly means that the problem will be *harder* to solve, not easier. Nevertheless, if a correlation can be established between sites of ancient tectonic activity and modern seismicity, an evaluation of seismic risk in intraplate regions can be made on a more fundamental basis than from historical seismicity alone.

Because nearly all of these ancient structures lie at depth beneath more youthful geologic strata, defining their nature and geometry relies heavily on expensive and time-consuming deep-sounding techniques. Perhaps the most complete investigation conducted so far in the eastern region has been in the New Madrid seismic zone, the site of the earthquakes of 1811 and 1812. There, aeromagnetic, gravity, seismic-reflection, and seismic-refraction data have outlined a northeast-trending crustal rift zone at depths of 5 km and greater beneath the modern surface of Arkansas, Missouri, Kentucky and Tennessee. This rift, thought to have been first

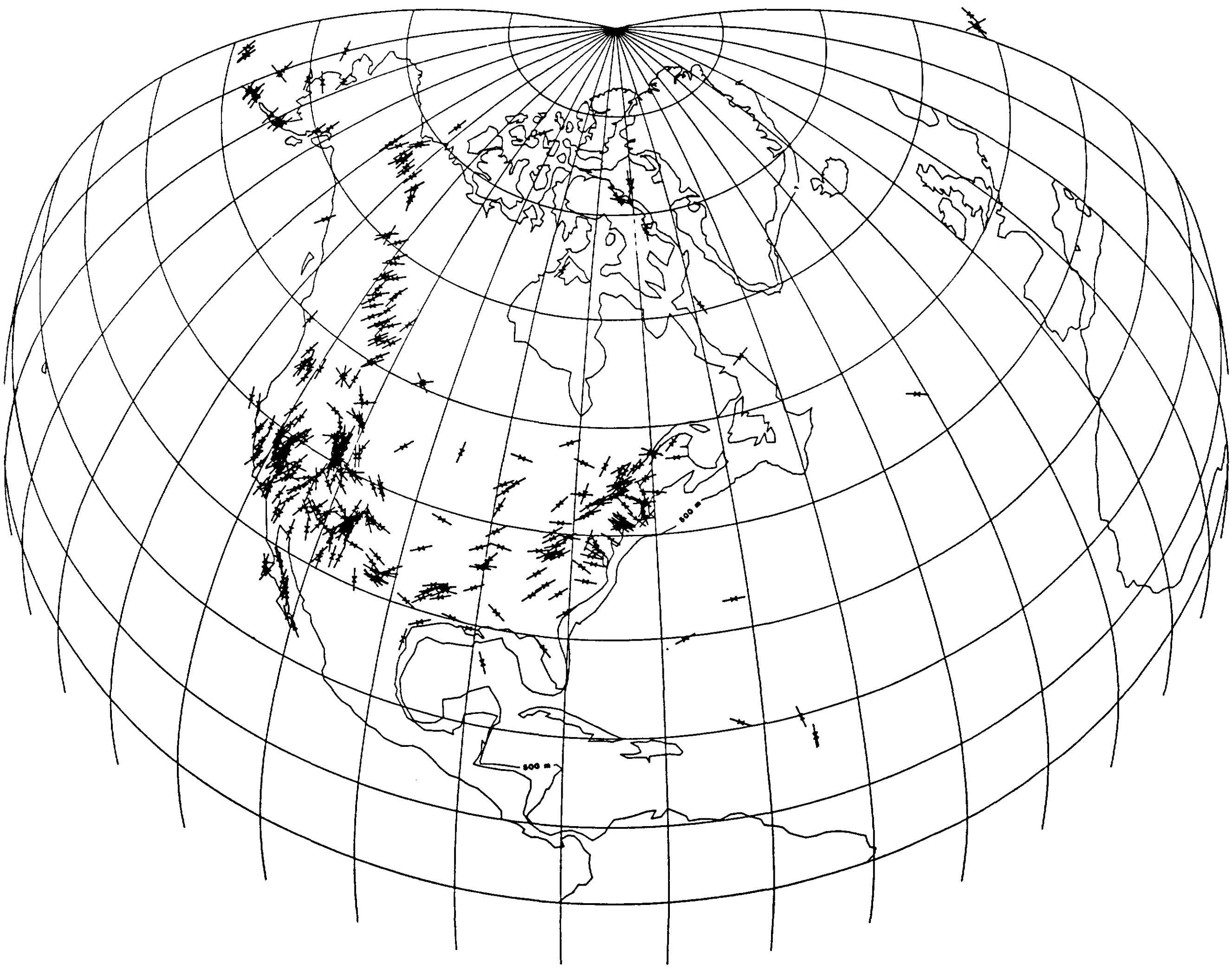


**Figure 19.** Orientations of the maximum principal horizontal stresses for the North American plate. Compiled by M.L. Zoback and M.E. Schiltz, from data of M.L. Zoback, M.D. Zoback, and M.E. Schiltz, "Index of Stress Data for the North American and Parts of the Pacific Plate" (U.S. Geological Survey Open-File Report 84–157), 1984.

activated in late Precambrian and (or) early Paleozoic time, is outlined by the heavy dark lines in figure 20, which also shows the contemporary seismicity to be closely associated with faults within the rift zone; these faults, inferred from seismic-reflection data, cut rocks of Paleozoic and Tertiary age. Present-day seismicity and faulting are the result of a WSW-ENE compressive stress field. Recurrent uplift of the Tiptonville dome and Ridgely ridge (northern and southern shaded areas, respectively, of figure 20) is apparently driven by faulting at greater depths, because basement highs underlie each. Along the Reelfoot scarp, which forms the east boundary of the Tiptonville dome, trenching reveals recent offset along a set of normal faults that is probably continuous with faults at depth.

Multidisciplinary investigations such as these (another is ongoing in the meizoseismal area of the 1886 Charleston, South Carolina, earthquake) have had an enormous impact on the assessment of the causes, effects, and attendant risks of earthquakes in the eastern United States. Mostly out of sheer ignorance as to why and where major earthquakes might occur in this region, it has ordinarily been necessary to assume fairly general and conservative (but almost certainly erroneous) propositions—for example, that such events could occur anywhere with comparable (albeit low) probability. The studies of the New Madrid seismic zone, however, reveal that the earthquakes are there for a reason, namely, the tectonic activity of a major, identifiable geologic structure. Such investigations, then, which are only beginning in the

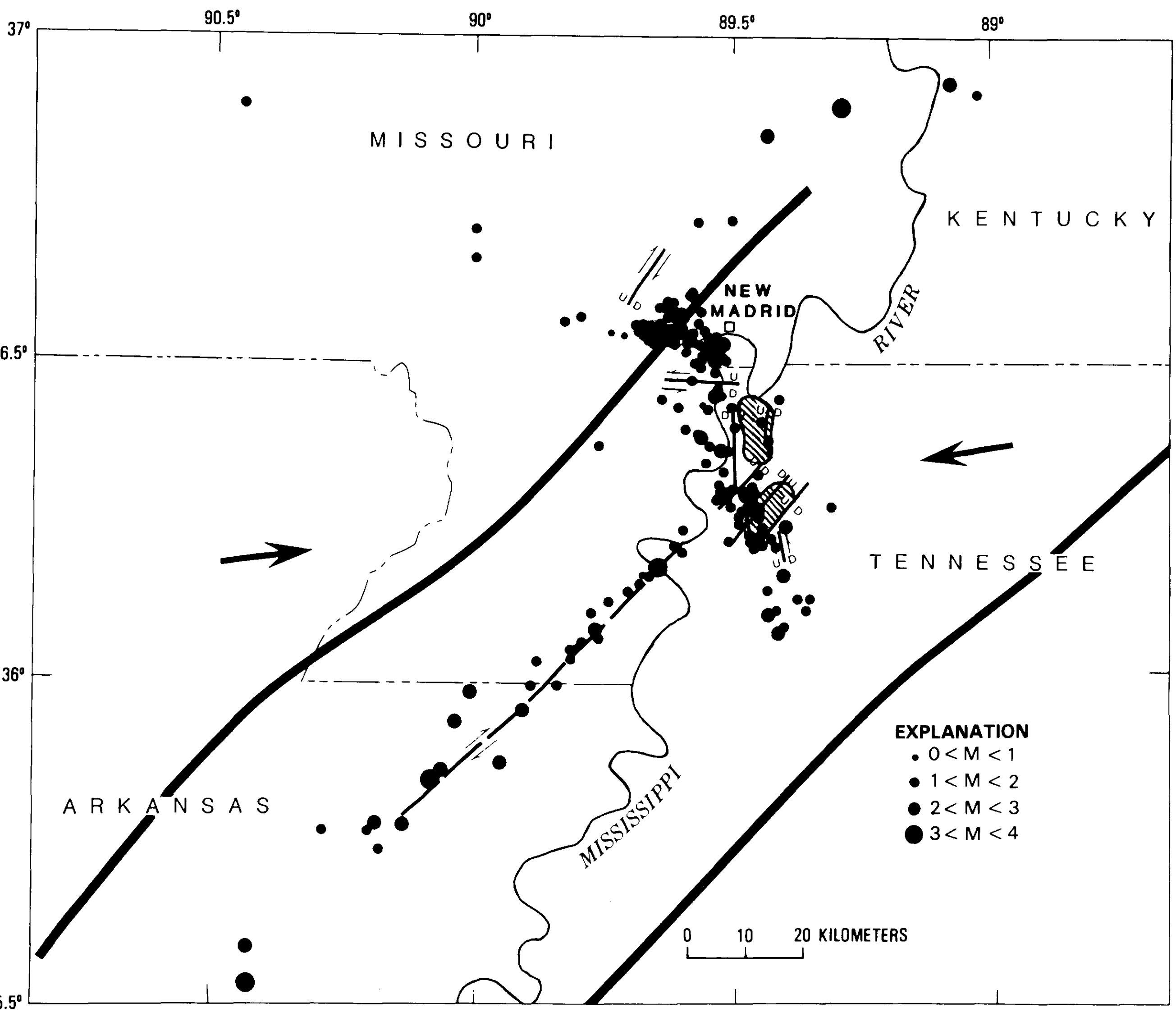



**Figure 20.** Seismicity and faulting in the New Madrid seismic zone. Dots, earthquake locations for the year 1979. Line segments are faults, with sense of motion indicated by opposed arrows and (or) **U** (up) and **D** (down). Bold lines, approximate boundaries of late Precambrian and (or) early Paleozoic rift zone; shading, currently elevated areas. From M.D. Zoback and M.L. Zoback, "State of Stress and Intraplate Earthquakes in the United States" (Science, v. 213, p. 96–104), 1981.

eastern region, have the potential of yielding vastly more accurate assessments of earthquake occurrence, hazards, and risk than have heretofore been possible.

Another important development, of significance to strong-ground-motion estimation in the eastern region, is the evidence from recently obtained, close-in seismologic recordings that eastern earthquakes (Arkansas, New Brunswick [Canada], New Hampshire, and South Carolina) are indistinguishable from California earthquakes at comparable $M_0$ or **M.** Figure 21 shows the Franklin Falls Dam accelerogram of the largest earthquake instrumentally recorded at close distance ($R \lesssim 10$ km) in the United States east of the Rocky Mountains, the Gaza, New Hampshire, earthquake of January 18, 1982 whose magnitude was 4.5. Also shown in figure 21 is an accelerogram of an aftershock (0350 G.m.t. Aug. 6, 1975; **M** = 4.8) of the Oroville, California, earthquake, written at much the same distance. Allowing for the fact that the Oroville aftershock was a slightly larger earthquake than the New Hampshire shock and thus wrote a slightly larger amplitude accelerogram at a fixed, close distance, these two accelerograms are very similar in overall amplitude, frequency content, and duration. This, in turn, implies that these two earthquakes share very similar overall dynamic properties.

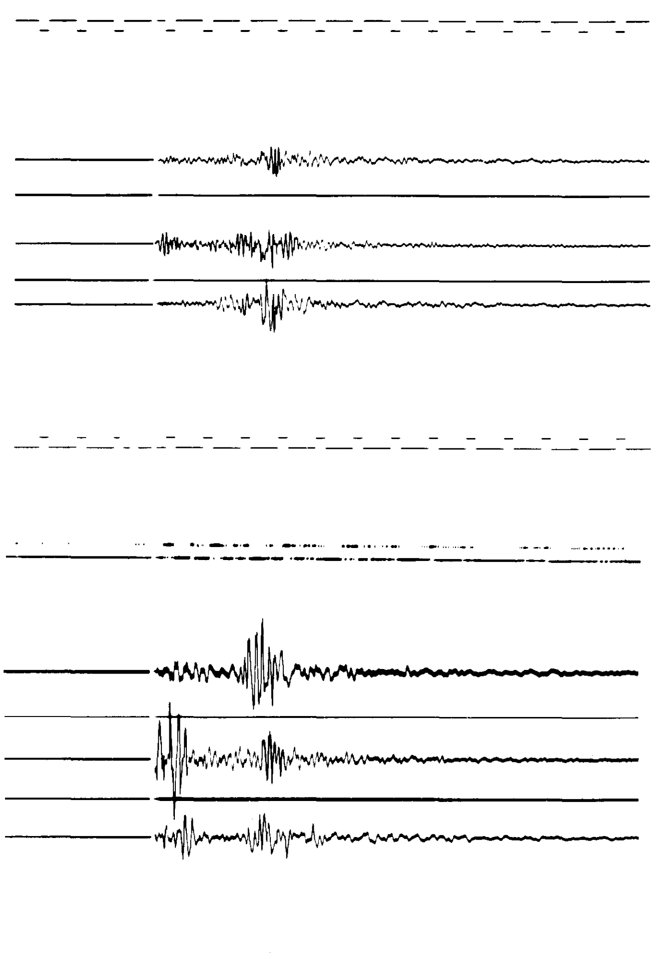


**Figure 21.** *Top,* Franklin Falls Dam accelerogram of the 1982 Gaza, New Hampshire, earthquake, with a peak acceleration of approximately 0.3 *g*. *Bottom,* Oroville Medical Center accelerogram of the Oroville aftershock of 0350 p.s.t. August 6, 1975, with a peak acceleration of approximately 0.5 *g*. On both accelerograms, the vertical component of motion is the center trace, with two orthogonal components of horizontal motion above and below it; bottom trace is a sequence of half-second time marks. Data of U.S. Geological Survey.

A more quantitative illustration of this similarity between other U.S. and California earthquakes is shown in figure 22. Here, seismic moments, source dimensions, and stress drops are given for six sets of earthquakes: the Arkansas swarm beginning on January 12, 1982; aftershocks of the New Brunswick earthquake of January 9, 1982; and members of the set of induced earthquakes at the Monticello Reservoir, South Carolina. The California earthquakes were recorded as aftershocks of the 1975 Oroville and 1980 Mammoth Lakes earthquakes, and near Anza along the San Jacinto fault zone. Above $M_0 = 2 \times 10^{19}$ dyne-cm (or $\mathbf{M} \geqslant 2.2$), all these events have essentially constant stress drops (1–10 MPa). More importantly, there is no way of distinguishing one set of earthquakes from another on the basis of stress drop. This is an important result because the constancy of earthquake stress differences figures prominently in high-frequency strong-ground-motion scaling relations, a subject we take up in the next section.

## Strong ground motion

Strong ground motion is the vehicle by which potential earthquake hazards become real ones, irrespective of whether the dynamically driven failure affects geologic structures, resulting in faulting or ground failure of a variety of types, or structures fabricated by man, resulting in some sort of damage, functional loss, or catastrophic collapse. Strong-ground-motion estimation thus plays a central role in earthquake-hazards reduction in general and is an essential ingredient of the earthquake-resistant design of critical structures in particular.

Although strong-motion seismology "began" 50 years ago, with the writing of four strong-motion accelerograms in the Los Angeles Basin by the 1933 Long Beach earthquake, strong-ground-motion-estimation studies have always been plagued by the uncertainties of having no observational constraints on how strong strong ground motion can really be. Although close-in records obtained for the 1940 Imperial Valley, 1966 Parkfield, 1971 San Fernando, and 1979 Imperial Valley earthquakes have been important milestones, they also set an ominous trend, as peak accelerations increased with each successive earthquake to 1.75 *g* for the recent Imperial Valley earthquake. Even at the present time, there are very few records at close distances for major earthquakes ($\mathbf{M} \geqslant 7$).

Since the San Fernando earthquake, however, and especially since 1975, great progress has been made in the field deployment of strong-motion accelerographs, both within the United States and abroad, commonly in the form of special arrays of closely sited instruments. Equally important have been developments in data-processing schemes, including laser-scanning and photodensitometer digitizing devices, whereby the increasingly larger data sets can be rapidly processed and made widely available. Finally, much progress has been made in the realistic seismologic modeling of strong ground motion, through the application of new source-modeling and wave-propagation techniques and better crustal-structure control.

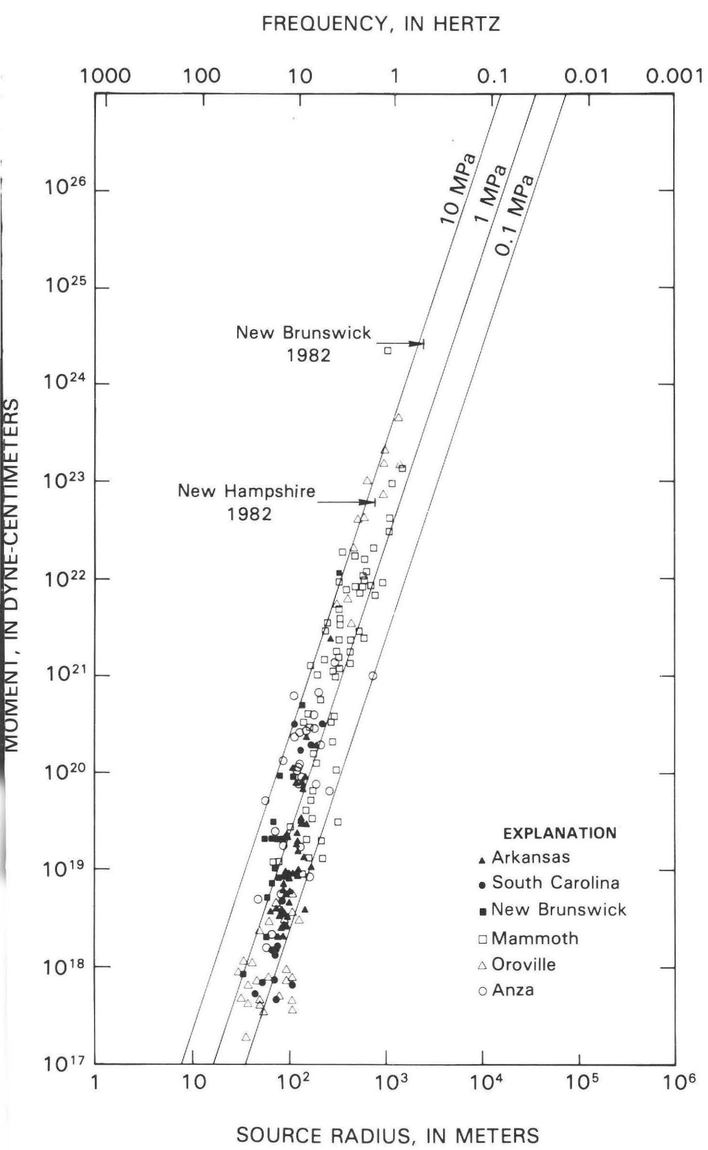



**Figure 22.** Seismic moments (ordinate), source dimensions (abscissa), and stress drops (labeled lines of slope +3) for six sets of earthquakes identified in lower right. There is no evidence for a population of eastern U.S. earthquakes (solid symbols) distinct from California earthquakes (open symbols). Modified, with addition of the New Brunswick and New Hampshire earthquakes, from L.C. Haar, J.B. Fletcher, and C.S. Mueller, "The 1982 Enola, Arkansas, Swarm and Scaling of Ground Motion in the Eastern United States" (Seismological Society of America Bulletin, v. 74, p. 2463–2482), 1984.

All these developments came together at the time of the 1979 Imperial Valley, California, earthquake. By then, the seismicity, tectonics, and structure of the Imperial Valley had been intensively studied, largely as a result of interest in the geothermal potential of the region. An extensive seismic-refraction experiment to refine the crustal structure of the Imperial Valley had been undertaken in January-March 1979, involving 40 detonations at seven shotpoints and 1,300 different recording locations. Finally, by 1979, the Imperial Valley had been densely instrumented with strong-motion accelerographs on both sides of the United States-Mexican Border, as a result of activities of the California Division of Mines and Geology, the California Institute of Technology, the U.S. Geological Survey, the Universidad Nacional Autónoma de México, and the University of California, San Diego.

Figure 23 shows transverse components of ground velocity obtained from 19 strong-motion accelerographs located in the immediate epicentral region of the earthquake. A total of 44 strong motion accelerographs recorded this earthquake in the United States, and 9 more accelerograms were written just south of the border in Baja California, Mexico. Of special seismologic interest are the records obtained from the 13-station El Centro array; the earthquake rupture passed between stations E07 and E06. About 1 km northeast of station E10, a remarkable set of records (not shown here) was obtained from the Imperial County Services Building, four reinforced-concrete columns of which partially collapsed during the earthquake. The 13 accelerograms written in this structure are of enormous engineering significance, in part because this is the first well-instrumented building to sustain significant earthquake damage, and in part because the timing and mechanism of structural damage can be inferred from the accelerograms.

Figure 24 (top) compares model calculations with the ground velocities recorded at stations E05, E06, E07, and E08. The model calculations take into account: (1) the finite area of the causative faults, both the Imperial (35 by 13 km) and the Brawley (10 by 8 km) faults; (2) the spatial variations of slip rate, rise time, rupture velocity, and orientation of the propagating rupture; and (3) the crustal structure of the area, as determined by refraction surveys, seismicity studies, and shallow drill logs. Figure 24 tells us that strong ground motion over a fairly wide frequency band can be realistically and deterministically calculated in advance for any source/site pair of interest, provided we can be specific in describing the source mechanism, crustal structure, and local site conditions. Figure 24 (bottom) illustrates components of displacement on the fault surface, determined from an inversion of the data in figure 23. Most of the faulting for this earthquake occurred at depths below 5 km.

At higher frequencies, important advances have been realized through the recognition that ground acceleration time histories closely approximate finite-duration, band-limited, white Gaussian noise during the strongest motion associated with the direct shear arrivals. The root-mean-square ground acceleration, $a_{rms}$, can be associated with a dynamic stress drop that is remarkably constant, at least for California earthquakes. In figure 25, this approach is used to model a large set of $M_0$-$M_L$ data for California earthquakes. The model calculations predict a value of $M_L$ for any value of $M_0$, given the above prescription for the shear waves and an $a_{rms}$ stress drop of 10 MPa for all events. Although the data themselves exhibit considerable scatter, much of which can be related to nonsource variables, on average it appears that the $a_{rms}$ stress drop of 100 bars is a stable and entire feature of California earthquakes, from the smallest that can be recorded in normal circumstances to the largest that have occurred. At present, this approach appears to provide a firm theoretical basis for calculating all manner of ground-motion amplitudes, including not only Wood-Anderson amplitudes (fig. 25) but also peak accelerations, peak velocities, and response-spectral ordinates, at least for $\mathbf{M} \cong 6\frac{1}{2}$. For larger

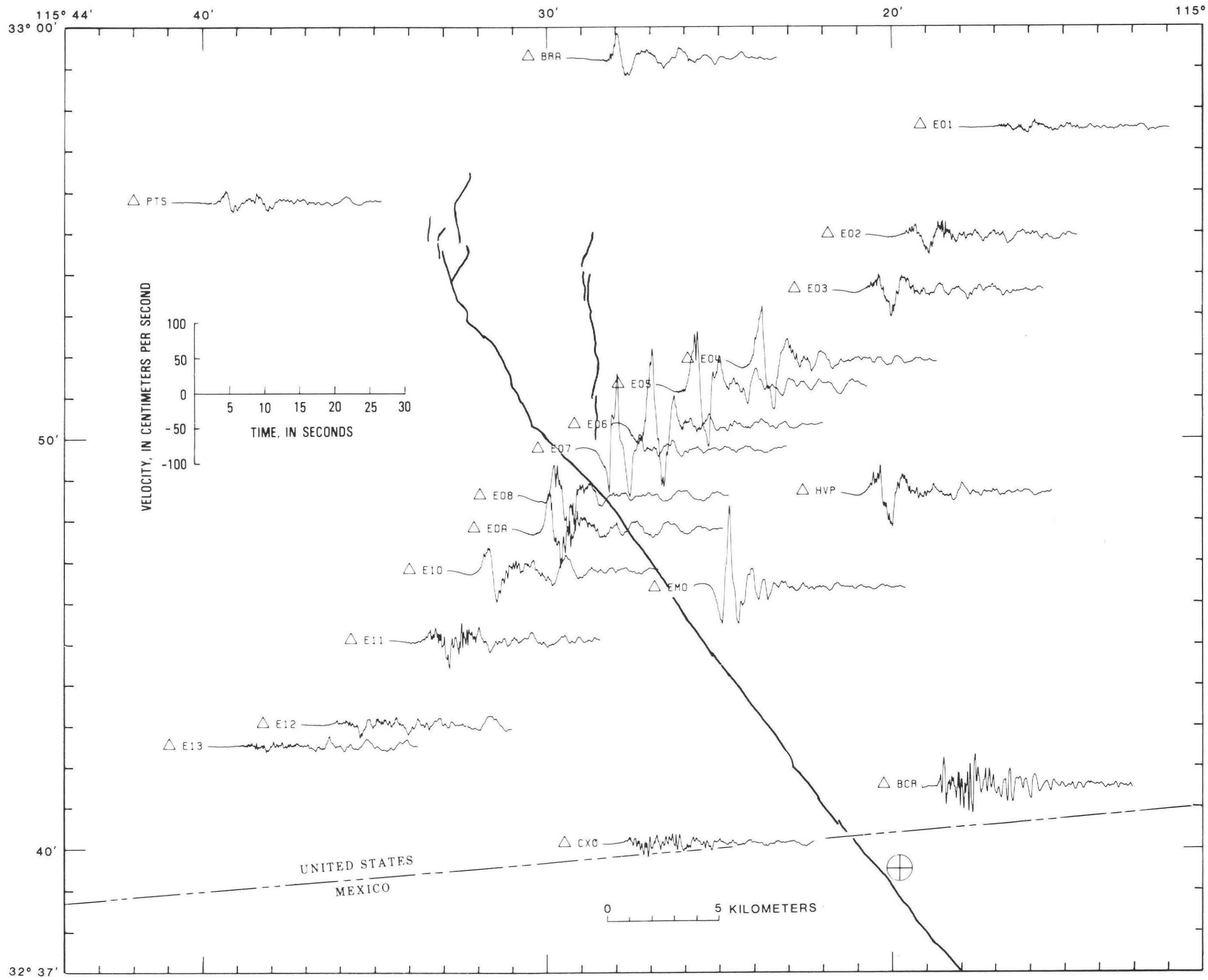



**Figure 23.** Ground velocity in the 230° direction, an azimuth approximately perpendicular to the Imperial fault, excited by the 1979 Imperial Valley earthquake. Most of the Imperial fault in this area and all of the Brawley fault (north-south fault segment in middle of figure) underwent surface rupture during this earthquake. Earthquake rupture initiated at location marked by circled cross, at a depth of 8 km, and propagated to the north. From R.J. Archuleta, "Analysis of Near-Source Static and Dynamic Measurements from the 1979 Imperial Valley Earthquake" (Seismological Society of America Bulletin, v. 72, p. 1927–1956), 1982.

**M,** some uncertainty yet remains, partly because the observational constraints are far fewer and partly because of the nature of large to great earthquakes, for which one faulting dimension (length) becomes progressively greater than its width.

Finally, as illustrated in the next section, important advances have been realized in the quantitative analysis of intensity data, assessments of shaking and damage that are vastly more numerous than instrumental measurements. For earthquakes antedating the relatively short instrumental era, such observations are the only direct information on earthquake effects (or even their occurrence). Moreover, intensity

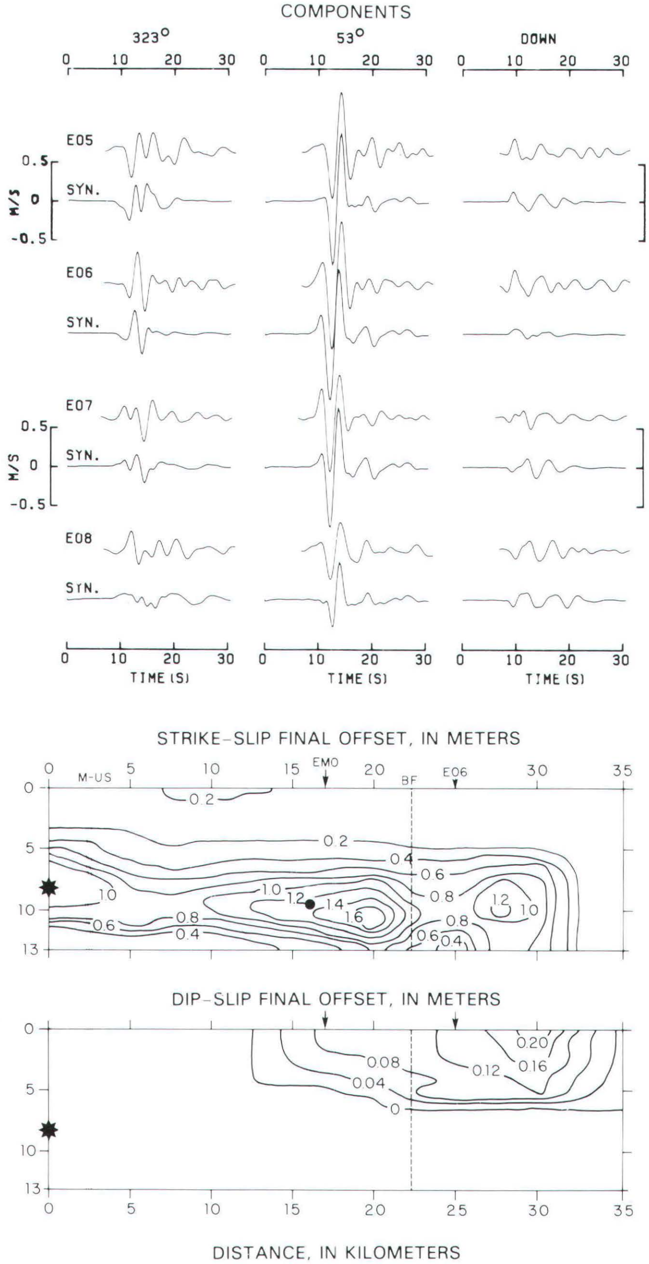



**Figure 24.** *Top,* observed and synthetic ground-velocity records of the 1979 Imperial Valley earthquake for three components of motion at sites E05, E06, E07, and E08 (fig. 23). *Bottom,* distribution of offset on the Imperial fault, obtained from an inversion of all the records in figure 23. From R.J. Archuleta, "A Faulting Model for the 1979 Imperial Valley Earthquake" (Journal of Geophysical Research, v. 89, p. 4559–4585), 1984.

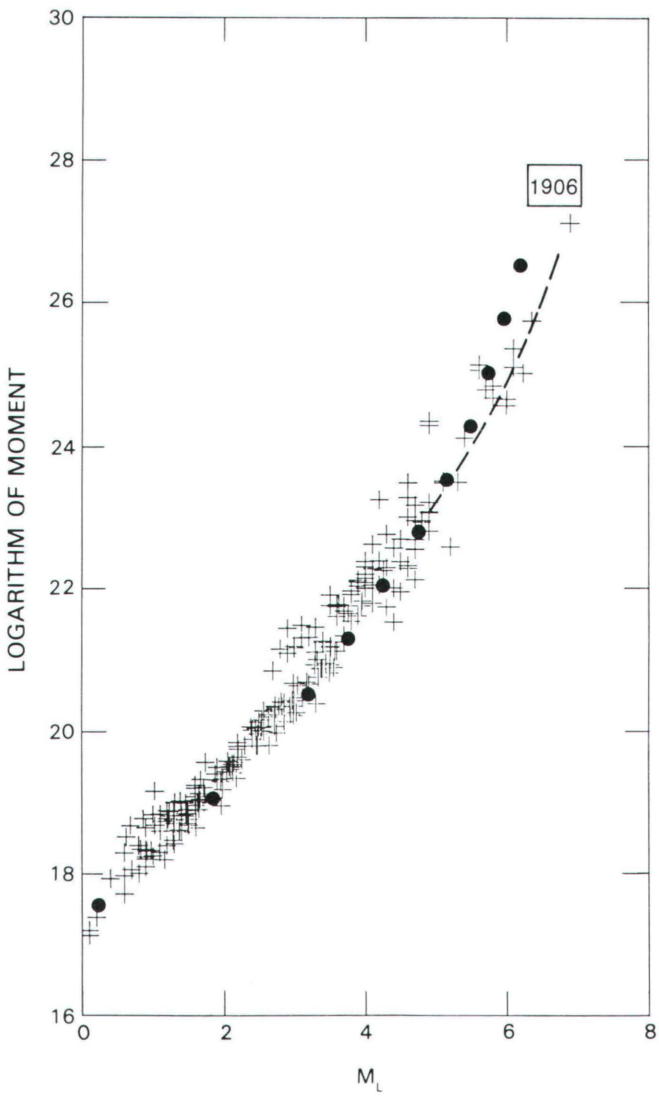



**Figure 25.** Moment-magnitude ($M_0$-$M_L$) data for central California earthquakes (crosses; box, 1906 San Francisco earthquake), and model calculations (dots) of $M_L$ given $M_0$, assuming that the maximum amplitude determining $M_L$ can be calculated on the basis that the shear-wave group at close distances is finite-duration, band-limited, white Gaussian noise. Dashed line represents a correction accounting for a slight discrepancy between the $M_L$'s determined at close and far distances. From T.C. Hanks and D.M. Boore, "Moment-Magnitude Relations in Theory and Practice" (Journal of Geophysical Research, v. 89, p. 6229–6235), 1984.

scales are closely tied to damage sustained by particular classes of structures. This association permits "damage functions" to be calculated for these structures, and so dollar losses can be estimated for some impending earthquake of specified size and location, given the building inventory within the area of strong ground shaking.

## Calculation of risk

In essence, earthquake-risk analysis is the calculation of the probability that something bad will happen as the result of one or more earthquakes during a specified time interval. In principle, that "something" can be almost anything—for example, structural damage, dam failure and flooding, or some form of ground failure—but risk calculations are fundamentally keyed to the probability that some level of ground motion $G$—a peak acceleration value, say, or some level of modified Mercalli intensity—will be exceeded (or not be exceeded) per unit time. For either a specific site or an entire region, then, one must integrate over all possible locations and sizes of earthquakes that can produce $G$ or greater within the time interval of interest. Figure 26 shows such a "risk map" for the 48 coterminous United States, here expressed as the horizontal ground velocity on hard rock (resulting from any earthquake) that, with a probability of 90 percent, will *not* be exceeded in 50 years. Such risk maps play an important role in fundamental design decisions, which must reckon the difficult compromises between a larger initial investment (for a more costly, more highly resistant structure) and the possibility of a larger economic loss some time later (to a less resistant structure). Such decisions, of course, are usually formalized in general-purpose building codes and in specific regulations for the construction of special-purpose structures, such as schools in California and nuclear reactors nationwide; risk maps such as figure 26 figure prominently in the rational development of these codes and regulations. These probability-of-exceedance calculations (of some $G$, at some place, for some time interval), moreover, form the basis for anticipated loss estimates, given that some relation between $G$ and the amount of damage to a particular structural form (say, wood-frame construction) is known. Such estimates of the replacement costs due to structural damage for repeats of the 1836 Hayward, 1857 Fort Tejon, 1906 San Francisco, and 1933 Long Beach, California, earthquakes have recently been calculated.

The calculations themselves, however, are a complex functional of various geophysical, geologic, and seismologic relations. These include $G$ expressed as a function of **M** and distance, definitions of seismic-source zones and capable

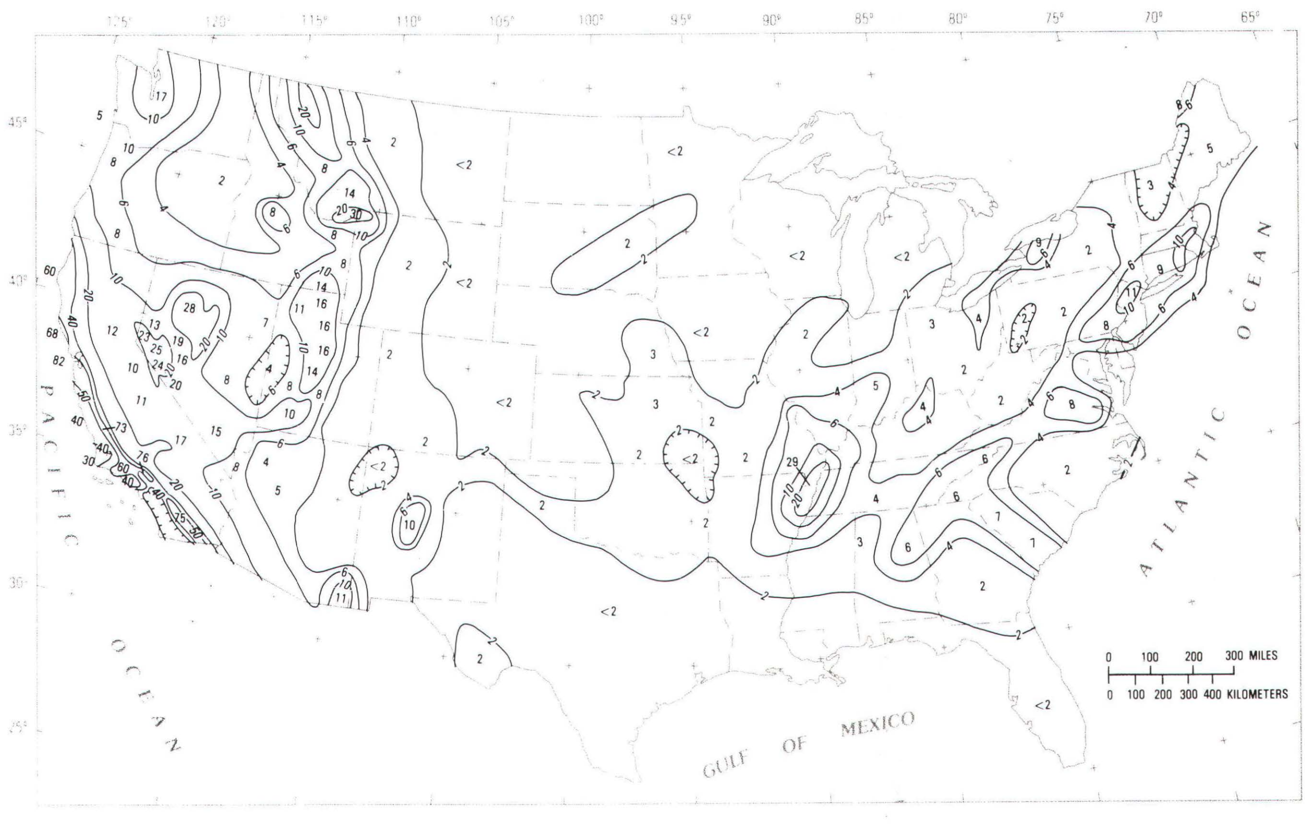



**Figure 26.** Horizontal velocities (in centimeters per second) on rock sites that have a 90-percent probability of not being exceeded in 50 years for the 48 coterminous United States. Plate 5 of S.T. Algermissen, D.M. Perkins, P.C. Thenhaus, S.L.Hanson, and B.L. Bender, "Probabilistic Estimates of Maximum Acceleration and Velocity in Rock in the Contiguous United States" (U.S. Geological Survey Open-File Report 82–1033), 1982.

faults, ground-motion-amplification factors for various geologic materials, and the local or regional earthquake-size/frequency-of-occurrence relations.

Any developments, several of which have been presented earlier, that lead to refinement of these relations plainly allow more accurate risk and loss estimates. The development of the characteristic-earthquake model has had especially important implications for the calculation of risk. The size-distribution/frequency-of-occurrence relation for earthquakes is empirically known to be of the form log $\mathbf{N} = a - \mathrm{b}\mathbf{M}$, where $\mathbf{N}$ is the number of earthquakes of magnitude $\geqslant \mathbf{M}$ within a given region and time interval. This relation permits an arbitrarily large earthquake to occur arbitrarily infrequently, a possibility that is implausible geologically but one that has been difficult to rule out in view of the long period of time involved. As discussed in the subsection above entitled "Inhomogeneous Faulting," however, there is a growing body of evidence that $\mathbf{M}$ or $M_0$ of the maximum-size earthquake can be defined in terms of the coherent length of any potential causative fault. With the growing body of slip-rate data that are now becoming available for these same faults, as discussed above in the subsections entitled "Crustal Deformation" and "Late Quaternary Faulting Geology," we can estimate the times of future earthquakes, if the date of the last maximum-magnitude event is known. Figure 27 summarizes such predictions for California in terms of their 30-year cumulative probability of occurrence.

It is straightforward to translate the information in figure 27 into ground-shaking probabilities for a fixed period of time, say, 30 years. We might calculate intensities directly and, from them, expectable losses; or we could calculate the distribution of ground-motion parameters about these source regions. Figure 28 presents such calculations for the islands of Honshu and Shikoku, Japan, here expressed as the probabilities of being shaken with JMA (Japan Meteorological Agency) intensity V for 20-, 50-, 100-, and 200-year intervals. The intensity distributions in figure 28 were obtained from recently developed relations between areal distributions of specific intensity levels and $M_0$, summed over all possible source regions on and near Honshu and Shikoku.

Although figures 26 through 28 reflect very recent developments and can be expected to be improved upon significantly in the coming years as better and more complete scientific information becomes available, they are, as statements of potential earthquake hazards at regional scales, essentially end products of the scientific enterprise. It is virtually certain that, in at least some areas of figures 26 through 28, potential hazards will become real during the next few decades. Indeed, the damaging earthquake of September 13, 1984, on the island of Honshu occurred almost at the center of the 90- to 100- percent probability-of-occurrence distribution in figure 28($a$). The mitigation of these impending threats, however, must be achieved in other arenas, in engineering-design decisions, land-use planning, and earthquake preparedness at several political levels. Such decisions, planning, and preparedness nevertheless depend heavily on realistic assessments of earthquake hazards on smaller spatial dimensions.

## Seismic zonation

Seismic zonation is the identification and evaluation of potential earthquake hazards on local scales. Primary hazards are those related to fault rupture at the Earth's surface and to strong ground motion. These phenomena, themselves known to vary spatially, can also effect a variety of secondary hazards—those related to ground failure (landslides and liquefaction) and inundation potential (tsunami run-up and flooding due to dam failures), for example. The potential occurrence and severity of these hazards, as well, can vary rapidly from one place to another.

In all such studies, the primary consideration is accurate mapping and determination of the rates of activity of the faults themselves, because these matters determine where the earthquakes will occur, how big they will be when they occur, how frequently such events will occur, and the distances to which any particular effect—at some fixed source strength—can occur. In addition, it is obvious enough—now—that manmade structures astride active crustal fault zones constitute especially dangerous situations. In California, this danger is recognized in the Alquist-Priolo Special Studies Zones Act of 1972, which requires special scientific investigations to be undertaken before building structures for human occupancy in the immediate vicinity of potentially active faults. These Special Studies Zones are detailed on official maps, with USGS 7½-minute topographic quadrangle maps as the base.

Comprehensive seismic-zonation studies in the United States began with a study of potential earthquake hazards in the 7,400-square-mile, nine-county San Francisco Bay region. These studies were first brought together in a single volume in 1975, which included contributions on faults and future earthquakes, bedrock-ground-motion estimation, discrimination of surficial sedimentary deposits, seismic response of local geologic units, liquefaction potential, landslide potential, and the predicted geologic effects of an $\mathbf{M} = 6.5$ earthquake on the San Andreas fault, centered behind Stanford University. From these results and those of more recent studies, figure 29 is drawn from digital data banks available to all local governments in the nine-county bay region. It shows the severity of ground shaking, measured in terms of the intensity scale describing the 1906 San Francisco earthquake, that may be expected as the result of major earthquakes along the principal faults in the Bay region.

Just as in the case of earthquake-risk estimation, seismic-zonation maps are only as good as the geologic, geophysical, and seismologic observations and techniques on which they are based. Since 1975, such investigations have become much more sophisticated, and their scope has broadened considerably, almost entirely as a result of advances since the inception of NEHRP. Many of these advances have been recounted earlier, but in addition to them, considerable progress has been made in the identification of liquefiable deposits and in the estimation of local ground-motion-amplification factors as a function of the material properties of near-surface sedimentary deposits. Thus, a nearly completed seismic-zonation study of the Los Angeles metropolitan area includes 14 contributions and some 1,200 manuscript pages, with very detailed predictions of various geologic effects and estimates of attendant economic losses. Essential studies for similarly comprehensive volumes are now being conducted in the regions in and around Anchorage, Alaska; Reno, Nevada; Salt Lake City, Utah; and Seattle, Washington.

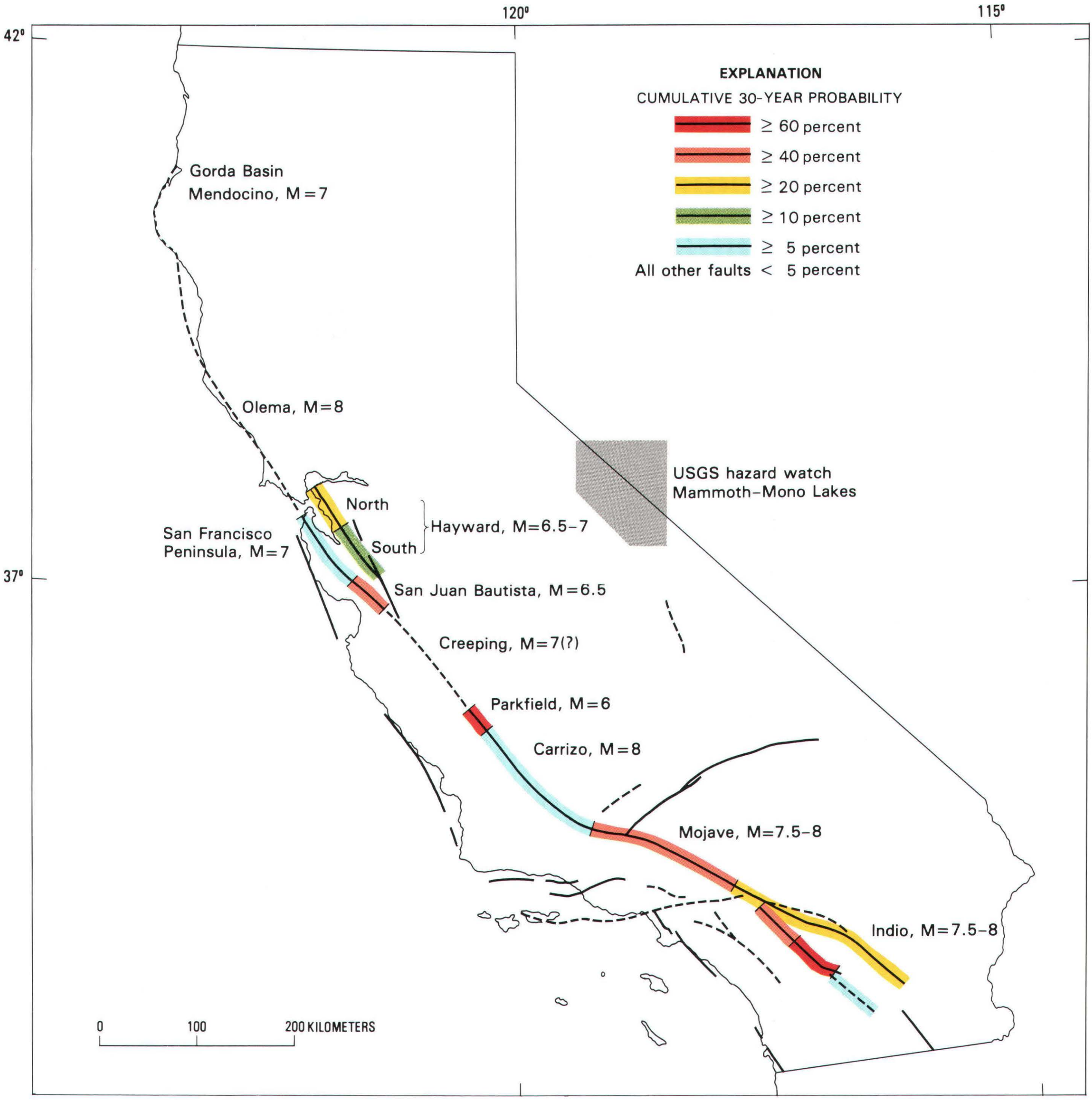



**Figure 27.** 30-year cumulative probabilities of occurrence of earthquakes along selected fault segments of the San Andreas fault system. Reconstructed from A.G. Lindh, "Preliminary Assessment of Long-Term Probabilities for Large Earthquakes Along Selected Fault Segments of the San Andreas Fault System in California" (U.S. Geological Survey Open-File Report 83–63), 1983.

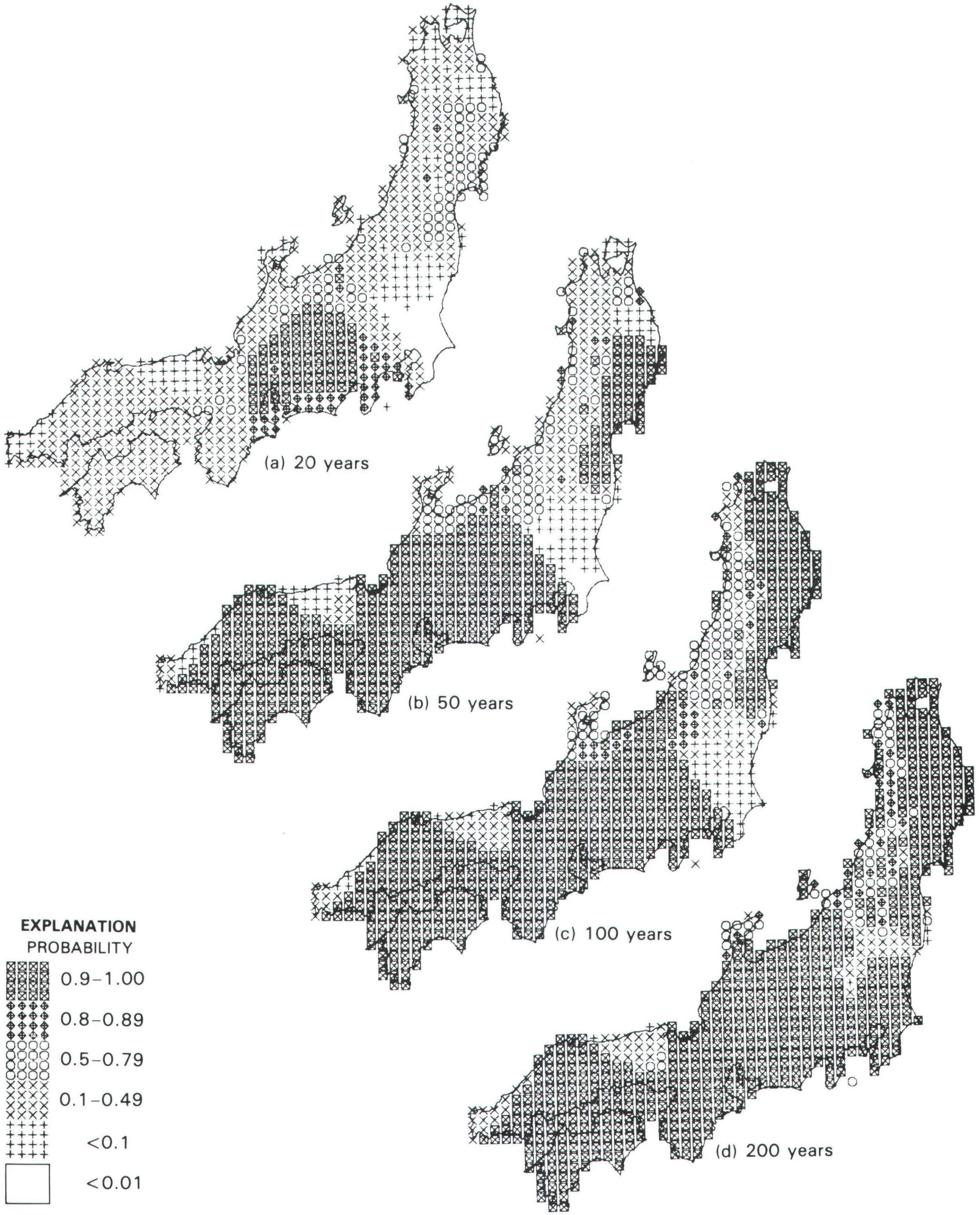



**Figure 28.** Probabilities that either interplate or intraplate earthquakes will give rise to ground shaking of JMA intensity ≥ V on the islands of Honshu and Shikoku, Japan, during the next 20 *(a)*, 50 *(b)*, 100 *(c)*, and 200 *(d)* years. From S.G. Wesnousky, C.H. Scholz, K. Shimizaki, and T. Matsuda, "Integration of Geological and Seismological Data for the Analysis of Seismic Hazard: A Case Study of Japan" (Seismological Society of America Bulletin, v. 74, p. 687–708), 1984.

Although seismic-zonation techniques are most generally applicable to the development of land-use policies, they are most intensively used in the siting of nuclear reactors. At the site itself, considerable effort is directed to searching for recently active faults and to determining local ground-motion-amplification factors. Consideration of the regional geology, seismology, and tectonics is used to define two levels of earthquake excitation, the Operating-Basis earthquake and the Safe-Shutdown earthquake, to which the plant must be designed. Inadequate or incomplete investigations in advance of construction can be far more costly than the studies themselves, as has been the case for the Diablo

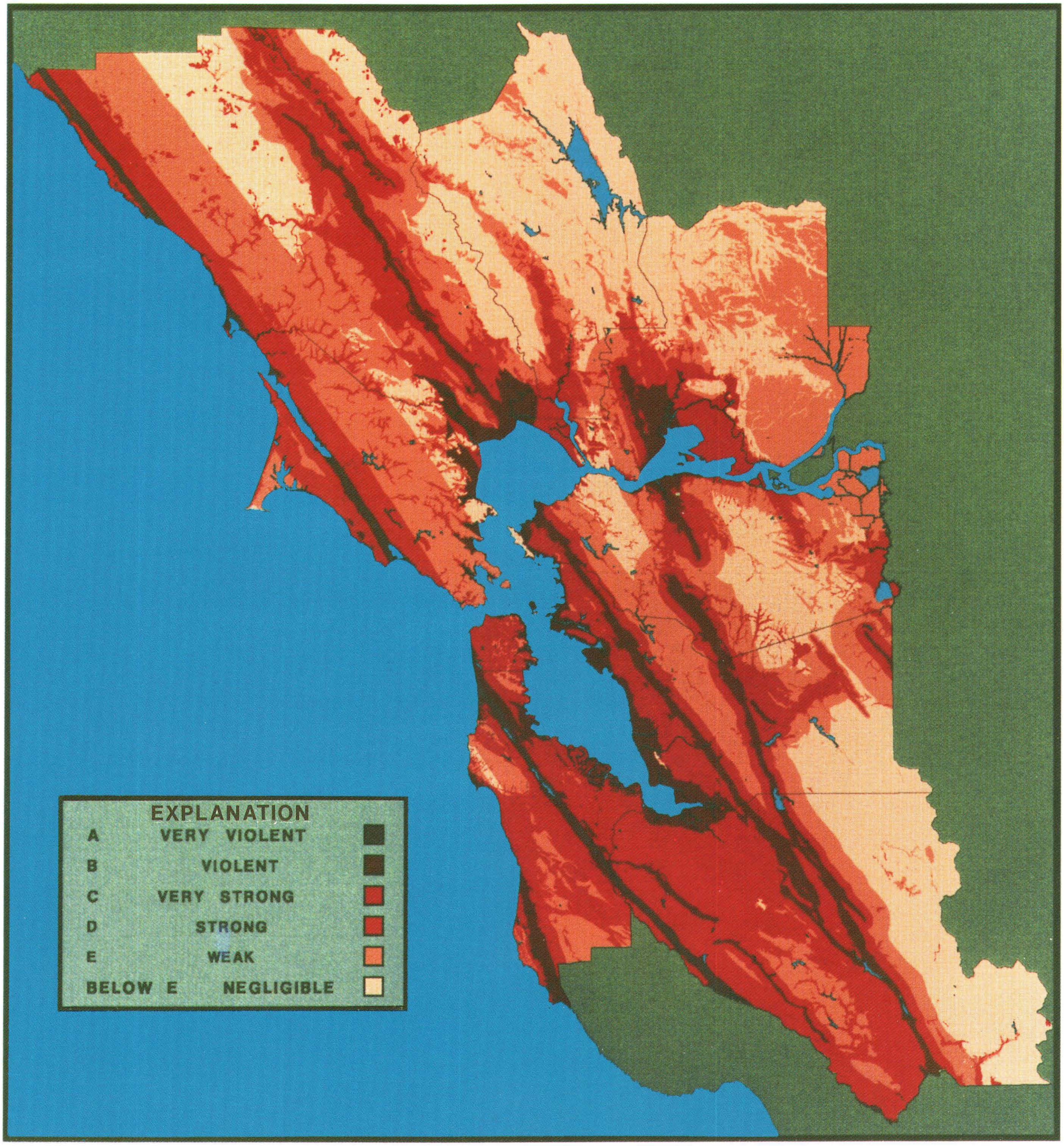



**Figure 29.** Maximum intensity of ground shaking expected for any earthquake in the nine-county San Francisco Bay region, according to the Bay Area Special Information System (BASIS).

Canyon, California, nuclear reactors. In this situation, the discovery of the Hosgri fault just offshore of Diablo Canyon, as a result of commercial oil-exploration studies, led to enormous expenditures and delays, in rationalizing the construction as designed previous to this discovery and in retrofitting the structure to higher levels of earthquake resistance.

## SCIENTIFIC PROGRESS AND PROGRAM PERSPECTIVES

The application of the scientific results discussed herein, as well as the many others that were not, to the meaningful reduction of earthquake hazards is a subject beyond the scope of this report. As intimated in the introduction, such an enterprise is much larger than the business of research science alone, with or without the equally important advances achieved in engineering disciplines not considered here. Even so, it would be by and large an empty exercise to conclude that NEHRP enjoys excellent scientific health without connecting, however briefly, the scientific accomplishments of the past 7 years to the objectives of the program.

Sure evidence that the Nation is progressing steadily toward the objectives of NEHRP is the widespread popular interest in matters earthquake, and much of this is attributable to the visibility of NEHRP itself. At least by the standards of pre-San Fernando earthquake times, this evidence is almost overwhelming. It exists in the form of how-the-Earth-works television movies, almost inevitably containing some footage on earthquakes; easy-to-read primers on earthquakes and their effects; homeowner's guides to do-it-yourself earthquake resistance; regular newspaper reporting of earthquake phenomena; and increasingly more readable accounts of ever more sophisticated concerns. Figure 30, for example, is a detail of the results portrayed in figure 29 for the San Francisco peninsula, a region including the subscription area of the *Peninsula Times Tribune*. Although such matters as these are rarely discussed in newspapers as they would be in scientific communications, in most cases this can be regarded as a positive development, not a negative one. There can be little doubt that the most certain protection against the dangers of earthquakes is an informed citizenry well prepared for them.

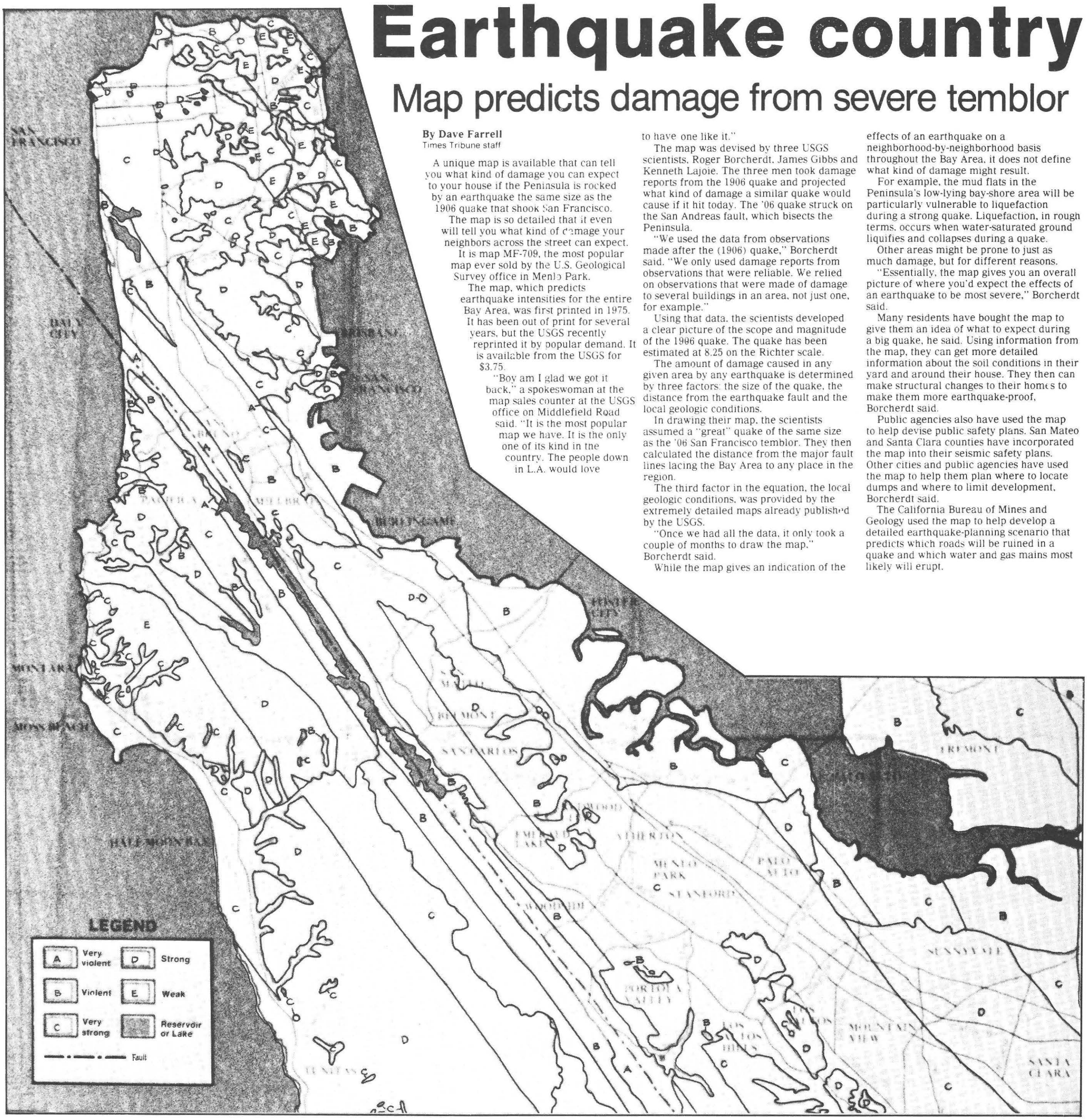
Just as importantly, this popular interest and concern have made themselves felt at city, county, State, and Federal political and public-policy levels. In California, for example, two major projects in earthquake-preparedness planning are now underway—the Southern California Earthquake Preparedness Project (SCEPP) and the Bay Area Earthquake Study (BAES). These efforts are funded by FEMA and State monies, and the Seismic Safety Commission, Office of Emergency Services, and Division of Mines and Geology of the State of California all participate in them. These projects also avail themselves of research expertise in university and government laboratories. In both cases, the project staffs work with representatives of cities, counties, businesses, schools, and various volunteer groups to stimulate and review earthquake-preparedness procedures. Specific model plans call for the development of a prediction warning system, response to earthquake predictions, response to unpredicted earthquakes, and recovery guidance.

But within the scientific context that is the basis of this document, how have the accomplishments recounted earlier led to progress toward the *scientific* objectives of NEHRP? We take these scientific objectives to be a full physical understanding of the cause and effect of earthquakes so as to (1) provide a complete identification and evaluation of potential earthquake hazards and (2) predict earthquakes, in whatever study areas are of concern.

It is commonly held, by citizens and scientists alike, that these matters of earthquake hazards and earthquake prediction are separate entities. Indeed, until quite recently, that part of the program administered by the USGS has operated on this basis, both internally and externally, since the inception of NEHRP. In the past several years, however, and as a direct result of scientific research conducted under NEHRP, many scientists now see this distinction in grayer tones. By example, the commonly used (nowadays) phrases "long-term prediction," "intermediate-term prediction," and "short-term prediction" plainly carry some implications about impending earthquake hazards on progressively shorter time scales. If we regard earthquake prediction (on whatever time scale) as simply a statement of impending earthquake hazards, we may also regard earthquake prediction as nothing more than one of many mitigation strategies. As such, the scientific essence of NEHRP is the accurate and complete evaluation of potential earthquake hazards and the mechanisms by which they become real. Here is a safe harbor for the scientists and engineers of NEHRP: it is widely agreed that the scientific and engineering advances achieved under NEHRP are now leading to—and will continue to lead to—meaningful reduction of the dangers attendant to earthquakes.

Although this logic properly underscores both the scientific significance and practical importance of long-term statements of impending earthquakes and the mitigation strategies timed to them, it misses the special nature of earthquake prediction, both as a unique mitigation strategy and as an endlessly fascinating research area. Even in such modern, industrialized, and affluent societies as the United States, for which it may be hoped that major loss of life can be averted through prudent land-use policies and earthquake-resistant design and construction, it may take a century or more for all of the hazardous structures in seismically prone areas to be razed or upgraded. When allowance is made for any structural degradation as a function of time, it is reasonable to anticipate that there will be a finite inventory of unsafe structures essentially forever. Earthquake prediction is the last line of defense against the building that might collapse in the event of a nearby earthquake; an accurate short-term prediction of that earthquake simply allows for the structure to be evacuated. As a special mitigation strategy, however, earthquake prediction only makes sense if the prediction is accurate within a narrow time window, measured in weeks or less. It is simply unrealistic to expect that all the unsafe structures in a major metropolitan area could be evacuated for a longer period of time.

Scientifically, short-term prediction has been an elusive goal. Although important and exciting results have been reported for a few earthquakes (and incompletely documented reports exist for a number of others), earthquakes with definitive precursory activity on time scales of months or less are very much the exception, not the rule. An impressive string of $\mathbf{M} \cong 6$ earthquakes have occurred in California




# Earthquake country

## Map predicts damage from severe temblor

By Dave Farrell
Times Tribune staff

A unique map is available that can tell you what kind of damage you can expect to your house if the Peninsula is rocked by an earthquake the same size as the 1906 quake that shook San Francisco.

The map is so detailed that it even will tell you what kind of damage your neighbors across the street can expect.

It is map MF-709, the most popular map ever sold by the U.S. Geological Survey office in Menlo Park.

The map, which predicts earthquake intensities for the entire Bay Area, was first printed in 1975. It has been out of print for several years, but the USGS recently reprinted it by popular demand. It is available from the USGS for $3.75.

"Boy am I glad we got it back," a spokeswoman at the map sales counter at the USGS office on Middlefield Road said. "It is the most popular map we have. It is the only one of its kind in the country. The people down in L.A. would love to have one like it."

The map was devised by three USGS scientists, Roger Borcherdt, James Gibbs and Kenneth Lajoie. The three men took damage reports from the 1906 quake and projected what kind of damage a similar quake would cause if it hit today. The '06 quake struck on the San Andreas fault, which bisects the Peninsula.

"We used the data from observations made after the (1906) quake," Borcherdt said. "We only used damage reports from observations that were reliable. We relied on observations that were made of damage to several buildings in an area, not just one, for example."

Using that data, the scientists developed a clear picture of the scope and magnitude of the 1906 quake. The quake has been estimated at 8.25 on the Richter scale.

The amount of damage caused in any given area by any earthquake is determined by three factors: the size of the quake, the distance from the earthquake fault and the local geologic conditions.

In drawing their map, the scientists assumed a "great" quake of the same size as the '06 San Francisco temblor. They then calculated the distance from the major fault lines lacing the Bay Area to any place in the region.

The third factor in the equation, the local geologic conditions, was provided by the extremely detailed maps already published by the USGS.

"Once we had all the data, it only took a couple of months to draw the map," Borcherdt said.

While the map gives an indication of the effects of an earthquake on a neighborhood-by-neighborhood basis throughout the Bay Area, it does not define what kind of damage might result.

For example, the mud flats in the Peninsula's low-lying bay-shore area will be particularly vulnerable to liquefaction during a strong quake. Liquefaction, in rough terms, occurs when water-saturated ground liquifies and collapses during a quake.

Other areas might be prone to just as much damage, but for different reasons.

"Essentially, the map gives you an overall picture of where you'd expect the effects of an earthquake to be most severe," Borcherdt said.

Many residents have bought the map to give them an idea of what to expect during a big quake, he said. Using information from the map, they can get more detailed information about the soil conditions in their yard and around their house. They then can make structural changes to their homes to make them more earthquake-proof, Borcherdt said.

Public agencies also have used the map to help devise public safety plans. San Mateo and Santa Clara counties have incorporated the map into their seismic safety plans. Other cities and public agencies have used the map to help them plan where to locate dumps and where to limit development, Borcherdt said.

The California Bureau of Mines and Geology used the map to help develop a detailed earthquake-planning scenario that predicts which roads will be ruined in a quake and which water and gas mains most likely will erupt.

**Figure 30.** Detail of figure 29, with accompanying text, as printed in the *Peninsula Times Tribune*, September 18, 1983 (Palo Alto, California).

since 1979, for example, including the 1979 Coyote Lake, 1979 Imperial Valley, 1980 Livermore, 1980 Mammoth Lakes, 1983 Coalinga, and 1984 Morgan Hill earthquakes, but none of these have been associated with definitive short-term precursors, even in an after-the-fact mode of operation.

From this, we might conclude that short-term earthquake prediction is now and forever beyond our reach. More rationally, we may conclude that we have simply performed the wrong experiments in the wrong places and at the wrong times. And, to be critical about matters, we might say that we have hardly conducted any experiments at all, to the extent we have allowed the occurrence of the earthquake itself to drive subsequent analysis on observations more often than not recorded from afar.

That is to say, the earthquakes we would most like to predict—and are most likely to predict—have not yet occurred, but we now know where to look, perhaps the single most significant accomplishment of NEHRP. In fact, an $\mathbf{M} \cong 6$ earthquake has already been predicted to occur near Parkfield within the next 5 years, and it is appropriate to execute experiments here with a very sharp focus. Also as a result of the scientific accomplishments of NEHRP, we now have a far better idea of what experiments to perform and how to perform them at a high degree of resolution. We know that we can obtain high-resolution seismologic and geodetic data when the appropriate instrumentation is clustered in dense arrays, and we also know that instruments in deep boreholes offer enormous advantages, both in signal-to-noise enhancement and in proximity to the source region.

With less certainty as to timing, many important developments also point to impending great earthquakes beneath the Shumagin Islands, Alaska, and along the Fort Tejon segment of the San Andreas fault that broke in 1857. The section of the San Jacinto fault near Anza has also been identified as a likely candidate for a $6 \lesssim \mathbf{M} \lesssim 7$ earthquake. And quite apart from the possible volcanic hazards, the Mammoth Lakes area is the most active region in California with respect to the occurrence of $\mathbf{M} \cong 6$ earthquakes.

The many accomplishments of NEHRP, then, have brought us to the edge of earthquake prediction on short time scales. They clearly point to the next step of intensively focusing on areas specifically identified as candidates for significant earthquakes, with experiments requiring substantial investments in intellect and funding. Until such experiments progress to their logical conclusion, we will not know whether short-term earthquake prediction is feasible or not. In the meantime, there is much to do.

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