DEPARTMENT OF THE INTERIOR

UNITED STATES GEOLOGICAL SURVEY

PROCEEDINGS OF THE NATIONAL EARTHQUAKE PREDICTION EVALUATION COUNCIL

JUNE 11-12, 1991

ALTA, UTAH

by

Virgil A. Frizzell, Jr.

Open-File Report 92-249

This report is preliminary and has not been edited or reviewed for conformity with U.S.G.S. publication standards.

1992

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PREFACE

The National Earthquake Prediction Evaluation Council (NEPEC) was established in 1979 pursuant to the Earthquake Hazards Reduction Act of 1977 to advise the Director of the United States Geological Survey (USGS) about issuing any formal predictions or other information pertinent to the potential for the occurrence of a significant earthquake. The Director of the USGS is responsible for deciding whether and/or when to issue predictions or other information pertinent to a prediction.

A prediction is defined as a statement on the time of occurrence, location, and magnitude of a future significant earthquake including an analysis of the uncertainty of those factors. NEPEC advises the Director concerning the completeness and scientific validity of the available data and on related matters. Duties include the evaluation of predictions made by other scientists, from within or outside of government, rather than issuance of predictions based on data gathered by NEPEC itself.

According to its charter, NEPEC, also referred to in this document as the Council, is comprised of a Chairman, Vice Chairman and from 8 to 12 other members appointed by the Director of the USGS. The Chairman may not be a USGS employee and at least one-half of the membership must be other than USGS employees.

NEPEC generally functions through the use of working groups organized by the USGS at the request of NEPEC. Working groups often include representatives from private industry, academia, and the USGS. Members of NEPEC who participate in a working group do not vote during NEPEC's evaluation of the results of the working group. After concluding its evaluation, NEPEC presents its recommendations to the Director, who bears ultimate responsibility for a decision concerning issuance of a prediction or other information.

The USGS has published the proceedings of previous NEPEC meetings as open-file reports; these reports are available from the USGS Open-File Distribution Center in Denver, Colorado.

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NATIONAL EARTHQUAKE PREDICTION EVALUATION COUNCIL
September 1991

Thomas V. McEvilly, Chairman
University of California

Robert L. Wesson, Vice-Chairman
USGS, Reston

Keiiti Aki
University of Southern California

William H. Bakun
USGS, Menlo Park

John N. Davies
University of Alaska

James F. Davis
California Division of Mines and Geology

James H. Dieterich
USGS, Menlo Park

Thomas H. Heaton
USGS, Pasadena

Arch C. Johnston
Memphis State University

Hiroo Kanamori
California Institute of Technology

William H. Prescott
USGS, Menlo Park

Kaye M. Shedlock
USGS, Golden

Joann Stock
Harvard University

Ray J. Weldon
University of Oregon

Virgil A. Frizzell, Jr., Executive Secretary
USGS, Reston

NATIONAL EARTHQUAKE PREDICTION EVALUATION COUNCIL

PROCEEDINGS OF THE MEETING OF JUNE 11-12, 1991

Alta, Utah

Council Members Present

Thomas McEvilly, Chairman	University of California
Robert Wesson, Vice-Chairman	USGS, Reston
William Bakun	USGS, Menlo Park
John Davies	University of Alaska
James Dieterich	USGS, Menlo Park
Thomas Heaton	USGS, Pasadena
Arch Johnston	Memphis State University
Hiroo Kanamori	California Institute of Technology
William Prescott	USGS, Menlo Park
Kaye Shedlock	USGS, Golden
Joann Stock	Harvard University

Virgil Frizzell, Executive Secretary, USGS, Reston

Participating Guests

M. Lee Allison	Utah Geological Survey
Walter J. Arabasz	University of Utah
Ronald L. Bruhn	University of Utah
Gary Johnson	Federal Emergency Management Agency
John O. Langbein	USGS, Menlo Park
William R. Lund	Utah Geological Survey
Michael N. Machette	USGS, Golden
Stuart P. Nishenko	USGS, Golden
David P. Schwartz	USGS, Menlo Park
Robert B. Smith	University of Utah

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JUNE 11, 1991
Morning Session

T.McEVILLY, Chairman of the National Earthquake Prediction Evaluation Council (NEPEC), opened the Council meeting by asking members, participants, and guests, to introduce themselves and by outlining the meeting's agenda (see Appendices A and B). All Members were in attendance except K. Aki, J. Davis, and R. Weldon.

R.WESSON, Vice-Chairman of NEPEC, presented an overview of Council activities with an emphasis on the transition from the Chairmanship of Lynn Sykes to that of McEvilly. At the termination of Sykes' tenure, NEPEC had completed a probabilistic assessment of the San Francisco Bay Region and had visited a number of areas of the country (Northern California, Southern California, the Pacific Northwest, and Alaska) that had been perceived as having a level of hazard that warranted attention. NEPEC had intended to visit the Wasatch area, but the press of business in California precluded such a visit.

At the outset of McEvilly's tenure, NEPEC prioritized areas needing attention, and the Wasatch area remained a high priority region. The present meeting at Alta, Utah, was delayed because of the need to reevaluate the San Francisco Bay Region in light of the Loma Prieta earthquake of 1989 as well as the need to address the so-called "Browning prediction."

T.McEVILLY and **R.WESSON** agreed that several options were available and suggested that Council Members and guests consider various options during the day's presentation and what sort of document might be used to present NEPEC's response to the Wasatch front, as well as options that would allow NEPEC to help focus attention on the other issues under consideration.

R.WESSON also noted that Randall Updike completed his term as Executive Secretary of NEPEC and that an open-file report entitled "1990 Proceedings of the National Earthquake Prediction Evaluation Council" (Updike, 1990) resulted from his efforts. Several Members joined Wesson by congratulating Updike for a job well done.

R.SMITH of the University of Utah (UU) presented an overview of the seismotectonics, seismicity, and paleoseismicity of the Intermountain Seismic Belt (ISB) as a background to discussion of the Wasatch front of central Utah (Smith and Arabasz, 1991; See Appendix C). He described the ISB as a six state region with a linear zone of intraplate tectonism on faults dominated by normal deformation (App. C, fig. 1).

The 7.3 magnitude Borah Peak earthquake, which had scarps 3 to 3.5 m high, focused National attention on the importance of normal faulting earthquakes. The Hegben Lake, Montana, earthquake of

1955, demonstrated that active normal faults may be located along preexisting structures, in that case a Laramide thrust fault.

The structures at mouth of Little Cottonwood Canyon, the location of the current meeting, were postulated by G.K. Gilbert to be responsible for the uplift of the Wasatch Range. Those same young structures trend through Salt Lake City. Large earthquakes occurring in populated areas along the Wasatch front would have catastrophic effects on the population and economy of the entire state, as well as throughout the surrounding six state region.

The ISB occupies a large part of the western U.S., forms a arcuate region 1000 km by 400 km, and is comprised of southern, central, and northern portions (App. C, fig. 1). Earthquakes are generally shallow, occurring less than about 20 km in depth, and are diffusely located without obvious correlation to mapped Quaternary or Holocene faults. The earthquakes are the result of differential strain between a more stable part of the North America plate and the Great Basin.

Since with the Borah Peak earthquake of 1983, much has been learned about structures responsible for normal faulting events in the region. Aftershocks were not located along the scarps, but instead on a zone displaced 15 km to the southwest (App. C, figs. 10, 11), implying substantial dip to the fault. The main shock occurred in the same zone at about 16$\pm$4 km depth. Leveling data is compatible with a dipping planar structure.

Data from several other areas indicate that such structures become listric at depth. Our model for earthquake hazard analysis is one of planar structures with large events nucleating at crustal depths (App. C, fig. 21), so not only do we have to be aware of the possible effects that may extend to adjacent mountain blocks, but we must carefully consider hazards associated with structures that surface somewhat more distally from the population centers.

Quaternary faults, including six historic earthquakes have been scaled comparing Ms versus fault surface length and maximum surface displacement. The scaling law for maximum surface displacement is the same as that of Bonilla and others (1984) and DePolo (1990)(See App. C, refs). The second law, that of Ms versus surface length, is quite different.

If one compares the length of Quaternary faults and fault segments in the ISB to determine possible magnitudes that might be generated by movement on the faults, one finds that the Wasatch fault zone dominates with post-glacial earthquakes in the magnitude 7.2 to 7.3 range. If our maximum surface displacement scaling law is used, larger events are possible along the Wasatch, with magnitudes of 7.4 to 7.5. This presents an enigma in some areas which have large scarps with short surface length displacement. Smith emphasized that the detailed information needed to analyze faults in this manner is quite limited.

In summary, the model appears to be one of earthquakes nucleating at the brittle/ductile transition along planar structures that are connected to relatively shallow low angle detachments. The uniform pattern of coseismic strain must result from long-term interseismic strain.

W.ARABASZ of UU presented information on seismic hazards along the Wasatch front. Normalized seismicity rates of earthquake occurrence by area in the ISB is lower by a factor of 4 when compared to the plate boundary in California. The number of large historic earthquakes (over moment magnitude 6.5) is four, with one occurring in Utah. Less than 10 percent of the earthquakes in the Wasatch front area can be located with adequate focal depth.

The catalog for 1962 through 1990 includes some 12,000 events; independent main shocks of M_L 5.5 or greater for 1850 to 1986 total 14 (Fig. 1). East-west extension on normal faults predominates. The threshold magnitude for surface faulting throughout the region is M_L 6.0 to 6.5, and maximum magnitude appears to be about M_S 7.5 to 7.7. The historical record lacks large surface faulting earthquakes on the Wasatch fault, and there is a notable paucity of small instrumentally-located earthquakes on the Wasatch fault. Perhaps no earthquake of $M \geq 5.0$ has occurred on the fault since 1847. A weak correlation exists between background seismicity and mapped active faults.

Comparison of geologic structure with seismic data derived from portable arrays presents some insights, but prompts many questions. In many experiments, clustered seismicity cannot be found on major active fault planes, and seismicity is truncated in some regions by detachment surfaces. Earthquakes occur on discordant structures at depth; in some cases, background earthquakes occur in deeper plates.

Sources for the 15 largest historical earthquakes (up to about M_W 6.6) are arguably unknown. Earthquakes up to magnitude 6.5 can be expected to occur randomly throughout the main seismic belt.

Since 1974, reliable focal depths have been determined for only 485 earthquakes from a population of 6400 events. These well located events do not delineate the active parts of the Wasatch fault. Background activity is most abundant north and south of the segments of the Wasatch that have ruptured at least once in Holocene time.

Epicenter maps for the ISB (App. C, fig. 1, 12) exhibit a seismic gap in the central Wasatch front area, where a microseismicity gap is especially notable north of Salt Lake City. That gap extends

$M_L \geq 4.0$ 1850-1986
(solid = $M \geq 5.5$)

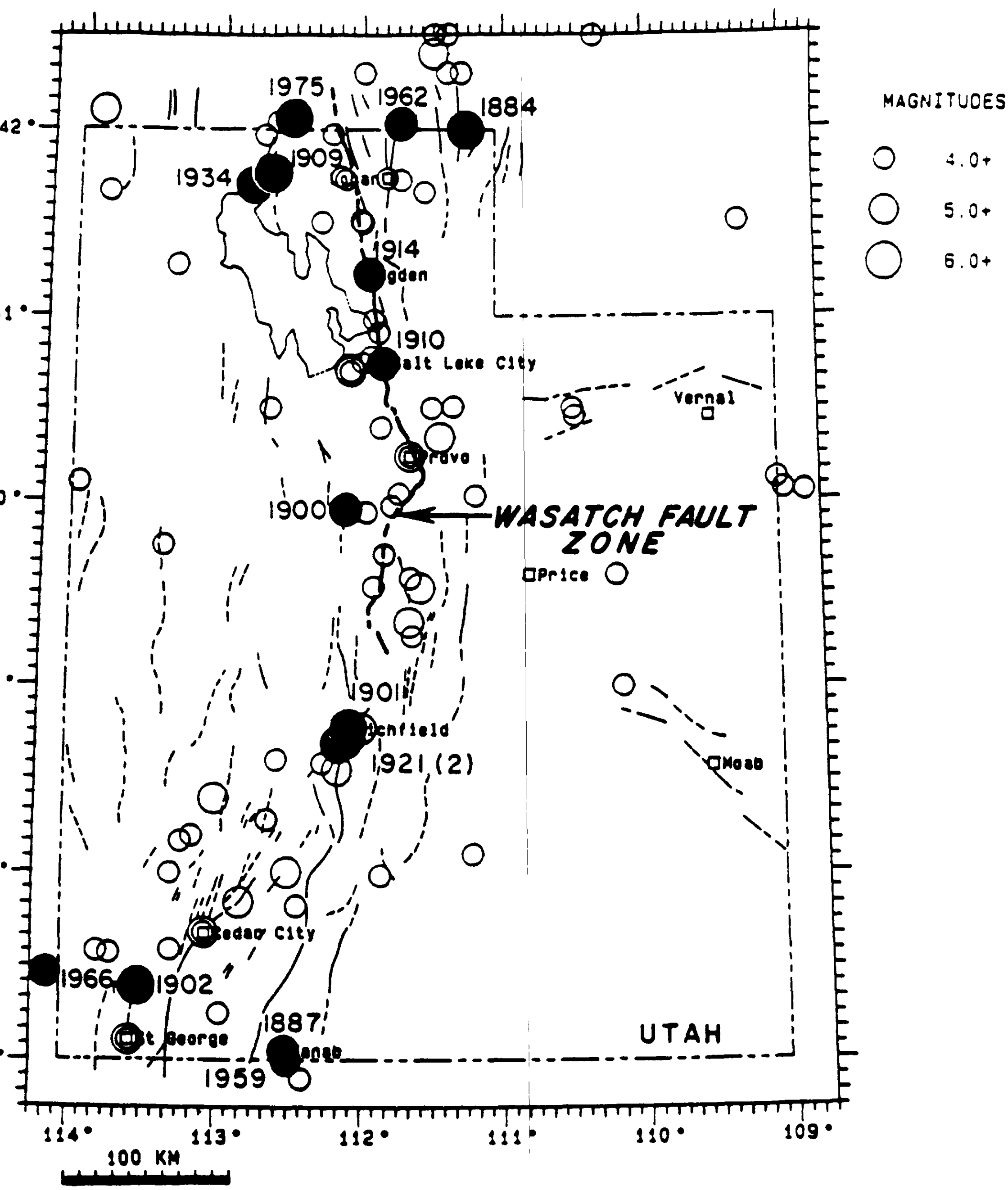

Figure 1: Epicenter map of the Utah region showing all independent main shocks of M_L 4.0 or greater (or Intensity V or greater), 1850 - 1986. Earthquakes of estimated M_L 5.5 or greater specially labeled. Data from University of Utah Seismograph Stations. From Arabasz and others (1987, fig. 6), which states that the base map of young faults in this figure is superseded by Arabasz and others (1987, fig. 5).

from about 39° to 41.5°, or from the southern end of the Levan segment to the northern end of the Brigham City segments (App. D, fig. 2), segments that have moved at least once in the Holocene.

In the main Wasatch front area, little hypocentral information is available. Not only do relatively few earthquakes occur, relative to neighboring areas, but, unlike other areas, there is little foreshock or aftershock activity related to the few main shocks.

For earthquakes along this active part of the Wasatch Front with magnitudes $\geq$ 3.0 from 1962 to 1990, for earthquakes $\geq$ 4.3 from 1938 to 1990, and for earthquakes $\geq$ 5.0 from 1900 to 1990, the hypothesis of a Poisson distribution cannot be rejected, Arabasz stated. No anomalous "trends" in seismicity have been detected and the background rate is lower than that north and south of the central active front in the ISB.

Cumulative plots of microearthquakes appear to hold some promise as a precursory tool. Cumulative plots of microseismicity within 50 km of the magnitude 6.0 Pocatello Valley event of 1975 exhibited an interesting pattern of changes in rates (Fig. 2A) Random distribution of microseismicity preceded a quiet period of 4.3 years. A precursory burst including a 4.2 magnitude earthquake and a clustering of events occurred a few years before the main earthquake. Random distribution of foreshocks immediately preceded the earthquake. This pattern has been exhibited by several other events in the region (Fig. 2B).

R.WESSON observed that south of about 41°, seismicity is mostly west of the trace of the main Wasatch fault and might be considered to be related to listric structures. North of about 41°, the seismicity is mostly east of the central portion of the fault, yielding a relatively persistent feature.

W.ARABASZ indicated that the events east of the Wasatch were diffusely scattered in that region, that their hypocenters were poorly resolved, and that the earthquakes may be related to lithospheric flexure. Maximum earthquake magnitudes range from 4.6 to 5.7, and maximum events that could occur there range from the high-6 to low-7 range. Some of the earthquakes can arguably be associated with structures at the surface.

M.MACHETTE stated that one difference about these structures is that the recurrence interval is 5 to 10 times longer than that of the Wasatch fault.

R.WESSON agreed that might be true from a geological point of view, but argued that from a seismological point of view the band appears to be virtually as active as the southern part of the Wasatch. He asked what fraction of the seismic hazard in Salt Lake City is associated with seismicity east of the Wasatch fault, what fraction is associated with activity west of the fault, and

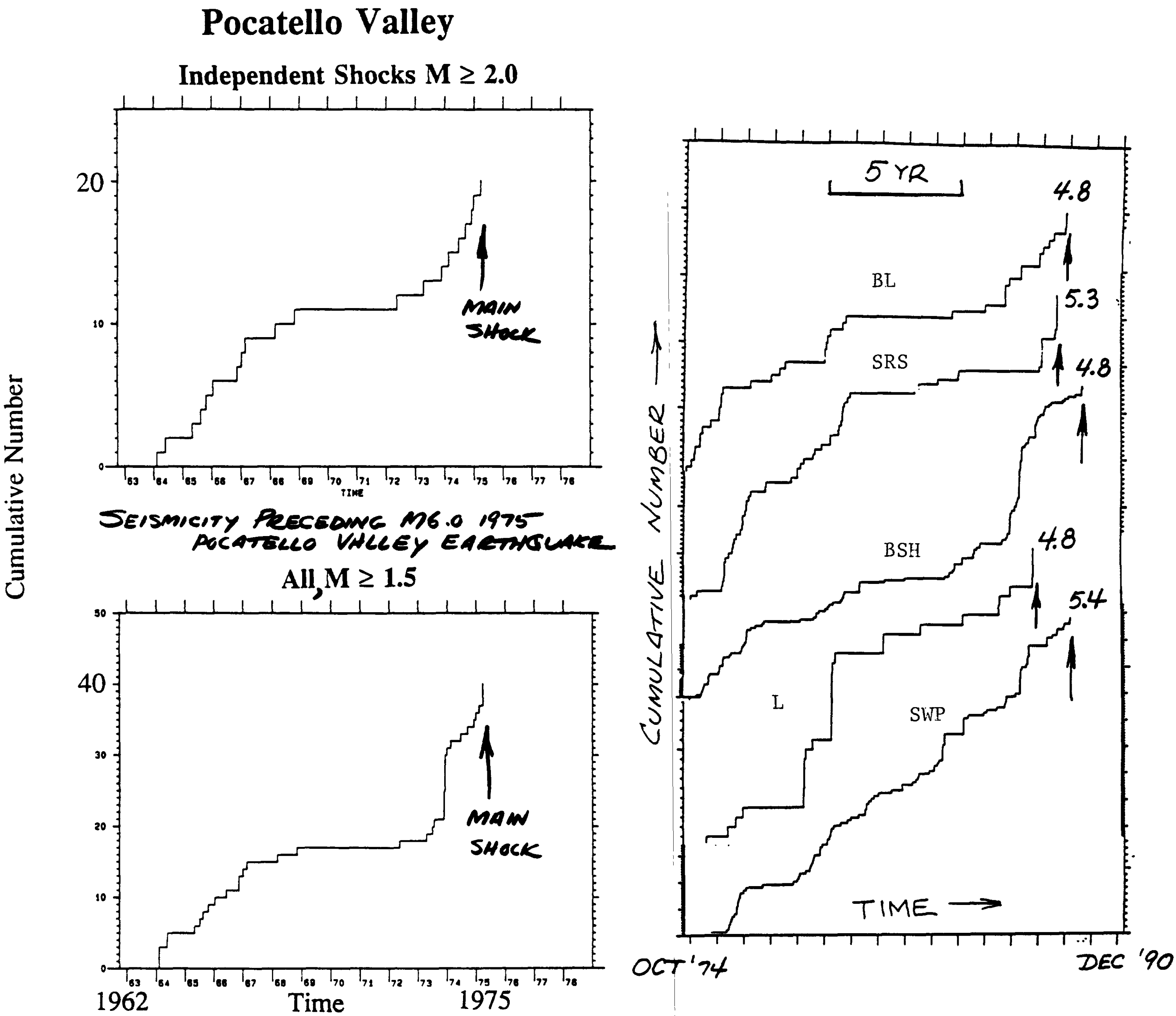

Figure 2: A. Cumulative plots of microearthquakes preceding the 1975 M 6.0 Pocatello Valley earthquake (W.J. Arabasz, written comm., 1991; see App. C, table 2, fig. 9 for more information and location). B. Cumulative plots of all earthquakes with magnitudes equal to or greater than M 1.5 within 25 km of respective epicenters prior to the 1988 M 4.8 Bear Lake (BL), 1987 M 5.3 San Rafael Swell (SRS), 1988 M4.8 Blue Springs (S), 1986 M 4.8 Lakeside (L), and 1988 M 5.4 South Wasatch Plateau (SWP) earthquakes (W.J. Arabasz, written and oral comm., 1991).

W.PRESCOTT agreed and stated that the rate was about a factor of 10 lower than California. Some of the lines look quite linear, but some have real problems, particularly with the 1972 survey. Prescott stated that although Savage would say that 9 of the points that fall off the trends are questionable at best, using any of three methods (using all data, throwing out the 9 questionable points, or throwing out the '72, '78, and '81 surveys) to reduce the data, the same rate, about 0.03 to 0.04 ± 0.01, can be derived for the net west of the fault.

One model applied by Savage (Fig. 3) assumes the Wasatch fault to be a normal fault with a 60^{o} dip that continues at depth, slipping on a surface with 60^{o} dip at about 15 km depth. Below about 20 km in an alternate model, the dip of the fault becomes much more shallow with slip occurring on the listric surface. The effect in the first model (Fig. 3B) creates very little strain west of the fault, but, in the second (Fig. 3C), significant extension is created west of the trace. Looking 50 km west and east of the fault, if 5 mm of slip is applied to the normal fault model, an amount required for 0.04 microstrain per year, areas to the east of the fault move towards the east (and are in extension) and areas west of the fault don't move much at all (and are in compression) and there is a slight dip into the fault. This model does not fit very well with extension. The alternate, listric, model provides fairly continuous distribution of velocity across the region and somewhat less vertical movement. The big difference is in the strain field which indicates extension across the region under study.

W.PRESCOTT also presented data from a level line run across the Wasatch fault by the National Geodetic Survey (NGS) and the USGS and analyzed by Wood and Vincent (1984). About 20 mm of valley-down change occurred during the period of 1959 (NGS leveling) to 1974 (USGS leveling). Little change occurred during the period of 1974 (USGS) to 1979 (USGS). From 1979 (USGS) to 1984 (NGS) about 10 mm of valley-down change occurred. Overall this yields 1 to 2 mm of change per year.

R.BRUHN of UU presented data on the geometry and state of stress on the Wasatch front. Investigations up and down the fault indicate that dips are mostly between 30^{o} and 50^{o}. Looking along strike of the Salt Lake segment, one gets the impression of a curtain with pleats and reentrants. Spurs or salients often end segments with subareal or buried bedrock ridges with more or less east-west trends. Microseismic activity appears to be concentrated along these segment boundaries, and is especially noticeable at the boundaries between the Nephi and Provo segments, the Provo and Salt Lake segments, and the Salt Lake and Weber segments. Many of the faults appear to have curved shapes, and this curvature may be due to segment curving as they terminate, not to any gross listric geometry.

whether there was a systematic difference in focal depths as the crust thickens to the east.

W.ARABASZ estimated that for events with magnitude 6.5 and greater, for a 50 year period, 50 percent of the hazard would be associated with the seismicity to the east, but, for a 250 year exposure period, the late Quaternary faults would dominate risk. The focal depth information that is available indicates depths ranging from 11 km on the north to 17 km on the south, with locations being a little deeper to the east.

T.HEATON asked that since virtually all the larger shocks have had foreshocks and most other Basin and Range normal-slip events have been preceded by clusters, were statistics available to estimate whether any given event was a foreshock to a larger event. He inquired whether an action plan had been developed for a 5.5 magnitude event occurring in Salt Lake City.

W.ARABASZ answered that given a magnitude 3 non-aftershock the likelihood of a magnitude 4 event or larger within 5 days and 10 km was about 2 percent. The suppressed microseismic activity makes such estimates difficult, but, if a moderate earthquake were to occur in the peripheral region, the possibility it was a foreshock would force a more rigorous analysis. No policy exists, but if a magnitude 3 event were to occur in the region, workers at UU would look for precursors, clustering, and for prior quiescence. If these phenomena had occurred, some probability scheme could be derived. Given a magnitude 5 in the Wasatch Front region, where seismicity is suppressed, however, no policy exists to determine whether a larger event would follow.

W.PRESCOTT presented a summary of the results of geodetic measurements made in the ISB for Jim Savage (USGS, Menlo Park). Strain accumulation observed over about the last 20 years in the western United States indicates that strain rates in the ISB are very low compared to other regions of the west.

A network across the range front fault was surveyed first in about 1972, but much of the net was west of the fault. In about 1980, the network was extended to the east across the East Cache fault. The deformation rate is so low that, even with 10 years of record to the east, conclusions would be premature. To the west, however, an extension of N85^{O}E is becoming apparent with rates on the order of 0.03 to 0.04 microstrain per year, and, although the data contain some noise and anomalies, Prescott noted that Savage has argued that strain is accumulating uniformly with time.

T.HEATON pointed out that the strain actually was higher than he would have thought, about 20 percent the rate on the San Andreas, about 0.2 microstrain per year.

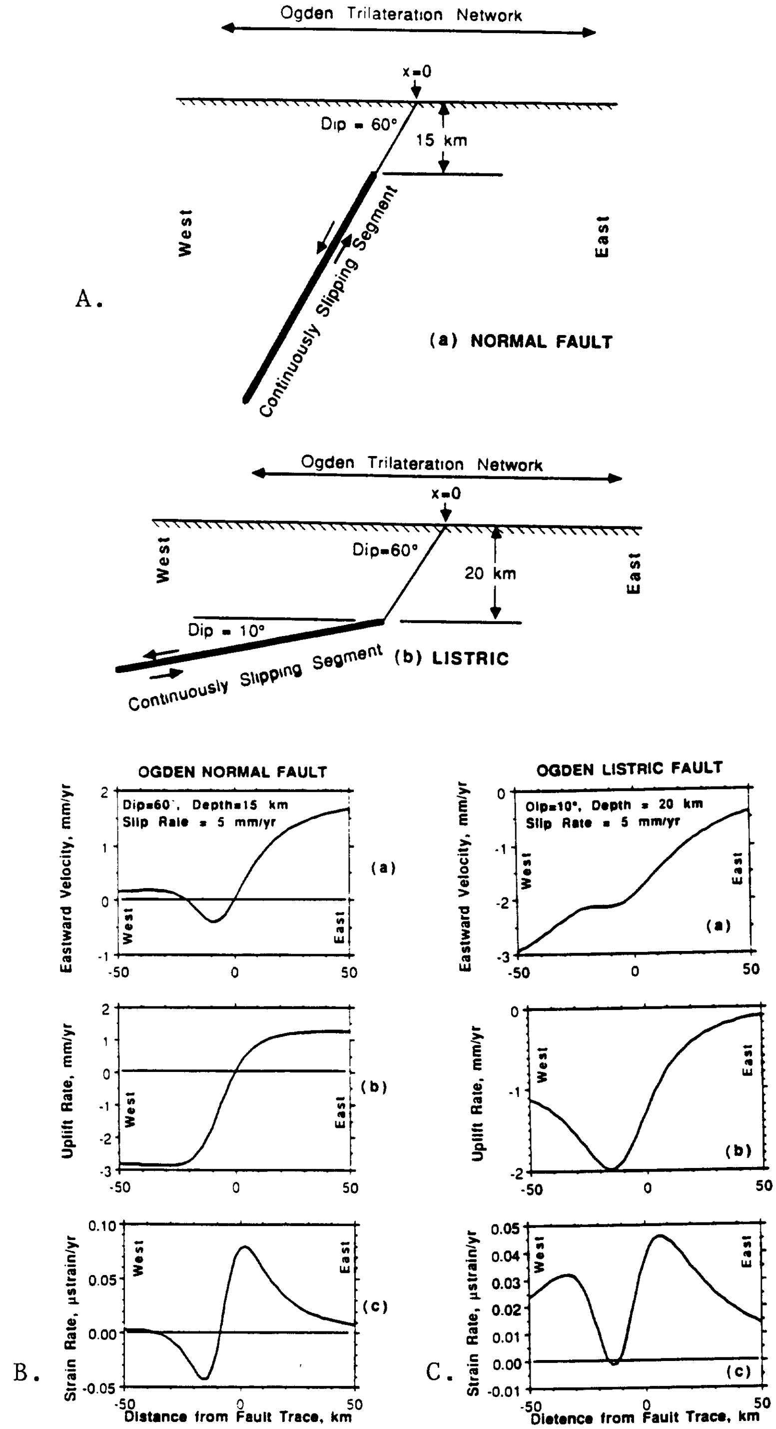

Figure 3. A. Two models of the Wasatch fault proposed to explain observed deformation of the Ogden network. Continuous uniform slip is postulated to occur on the deep segment of the fault (heavy line extending indefinitely down and to the left). B. Surface deformation predicted by the normal fault model of A(a) for a slip rate of 5 mm/yr on the creeping segment of the fault. C. Surface deformation predicted by the listric fault model of A(b) for a slip rate of 5 mm/yr on the creeping segment of the fault. From Savage and others (in review).

JUNE 11, 1991
Afternoon Session

M.MACHETTE with the USGS in Golden began a summary of Quaternary geologic studies in the region by describing the V-shaped wave of late Quaternary tectonics that points at the Yellowstone caldera (App. D, fig. 1; App. C, figs. 1, 23). Northwest trending faults north of the Snake River Plain have recurrence intervals ranging from 8,000 to 15,000 years and include several faults that have had movement more recently than about 15,000 years, including the 1983 Borah Peak rupture in central Idaho and the 1959 Hebgen Lake rupture in southwest Montana.

The belt of young faulting continues to the south and straddles the Colorado Plateau-Basin and Range provinces boundary. The Wasatch fault (Machette and others, 1991) is a prominent element of that structural boundary and is larger, longer, and has higher slip rates than other faults in the region. It is clearly capable of generating large earthquakes with magnitudes on the order of 7.0 to 7.5 and is comprised of ten segments that extend from southern Idaho to central Utah (App. D, fig. 2, table 1). The three northernmost segments and the southernmost segment do not appear to have been active in the Holocene.

The intervening segments (from south to north, the Levan, Nephi, Provo, Salt Lake City, Weber, and Brigham City segments) have ruptured in the Holocene, and the Nephi through Brigham City segments have all ruptured many times in the Holocene, at least three times in the past 6,000 years. These segments have about 2,000 year recurrence intervals, although each segment has its own rupture history and the recurrence intervals are not regular.

The landscape along the Wasatch front is underlain by a variety of materials with differing ages. With the exception of the Salt Lake City segment, which has mostly been covered by urbanization, many opportunities still exist to improve the chronology that has been developed over the past 10 years or so from about 50 trench and natural exposures.

The Wasatch fault presents a real hazard: It is the longest, most continuously active normal fault in the contiguous United States. It has had slip rates of 1 to 2 mm for the past 5,000 years, which are high for the extensional terrain of the Basin and Range. It is moderately seismic, but with no large historic events. Only two magnitude 5 events can reasonably be placed on it.

The timing of movement on segments during the past 6,000 years are presented in a diagram that depicts various patterns for different segments of the Wasatch (App. D, fig. 4). Six events have occurred along the fault in the last 1,100 years. Over the last 6,000 years, the average composite recurrence interval has been about 400 years. However, in the last 1,500 years, the interval has been about 220 years. Questions include whether there is a clustering of events in the last 1,500 years and whether there are periods of more random activity.

M.MACHETTE concluded with a diagram of age versus fault offset that illustrates a change in slip rate. A long term rate of 0.1 to 0.3 mm per year on the Wasatch fault has been determined using deposits 50,000 to 250,000 years old; whereas the rate is 0.5 to 2.0 mm per year over the last 10,000 to 20,000 years, averaging about 1 mm for the post-Bonneville deposits. This 10-fold change in slip rate may be an artifact caused by changes in lake level.

W.LUND of the Utah Geological Survey (UGS) presented details from the Mapleton site to illustrate the importance of information derived from trenches for the development of fault histories. Although the Weber segment has changed length during its recent faulting history, the Weber and Brigham City segments have not moved in a single event. A good data base exists for large surface rupturing events on the central segments over the last 6,000 years. The Bureau of Reclamation has some evidence for a larger number of events on the Provo segment near the boundary with the Nephi segment; this may be due to an event on the Nephi segment that continues on past the segment boundary for a couple kilometers.

Details from trenches across the Provo segment indicate that the penultimate and ultimate events at the American Fork and Mapleton trench sites are essentially the same age. This allows older terminology which divided the Provo segment into shorter segments to be set aside. Thus, the number of segments has gone from 6 to 12, and now to 10.

D.SCHWARTZ with the USGS in Menlo Park presented the geologic information and conceptual recurrence framework that he and Stuart Nishenko are using to develop estimated probabilities for the Wasatch fault. For the Wasatch fault, a tremendous amount of data yields a pretty good idea about recurrence, which the can be addressed in two ways. One involves Poissonian probabilities, and the other involves conditional probabilities.

For the model yielding conditional probabilities, several assumptions concerning characteristic earthquakes, segmentation, and recurrence are made. The characteristic earthquake model is appropriate for the Wasatch fault; it was developed here. One sees repeated slip of the same amount for earthquakes at independent sites. Although the segmentation model is not perfect, it is fairly robust and its acceptance seems to be higher here than in California for the San Andreas. To present conditional probabilities, some assumptions about recurrence are required. Time dependency or linear strain accumulation seems to be required, and one needs to know the average recurrence interval and elapsed time since the last event.

A time-space plot (Fig. 4) forms the basis for the estimates. Based upon a reinterpretation of unpublished data collected 10 years ago, we have added an event on the Salt Lake segment about 10,000 years ago. A couple models can be postulated from this modified data set. Three events occurred at fairly regular intervals. The long-term slip rate on the Salt Lake segment is about 1 mm per year for the past 14,000 to 15,000 years. The slip per event is about 4.5 to 5 m, the largest along the entire fault. Thus, based on the paleoseismicity, a recurrence interval of a little more than 5000 years can be derived. The calculated rate is 4750±987 years. This suggests that some sort of time dependent behavior is producing quasi-periodic recurrence.

Three events also occurred on the Provo segment. Paleoseismic recurrence interval is 2000±440 years, and the calculated recurrence interval is 2200±220 years, based upon 2.5 m slip and 1.25 mm/year.

These rates are based upon one datum, the dates based on trenches, and the slip per event comes from the trenches. For these two segments, the real recurrence and the calculated recurrence are similar. Schwartz said that this information surprised him, because he had always considered the Wasatch to be fairly variable and random in its behavior.

Depending upon which events one includes, the Weber segment might have a paleoseismic interval of 1400 to 1200 years and a calculated interval of 1300 years. Perhaps quasi-periodic behavior best explains movement for this segment, although a cluster model also would be appropriate.

A cluster model appears to be appropriate for two segments. In such a model two events closely spaced in time would be followed by a period of quiescence followed by two more events. The Brigham City, with two events fairly close in time, and the Nephi, with two events postulated in the range of 4000 to 5000 and a three thousand year interval to about 400 years ago, segments exhibit such a pattern.

In summary, Schwartz said that basic data from the Wasatch allows derivation of estimates regarding the timing of the next events. The Salt Lake City, Provo, and, perhaps, Weber can be addressed using quasi-periodic recurrence models. For Brigham City, Weber, and Nephi, clustered and quasi-periodic activity models cannot be distinguished.

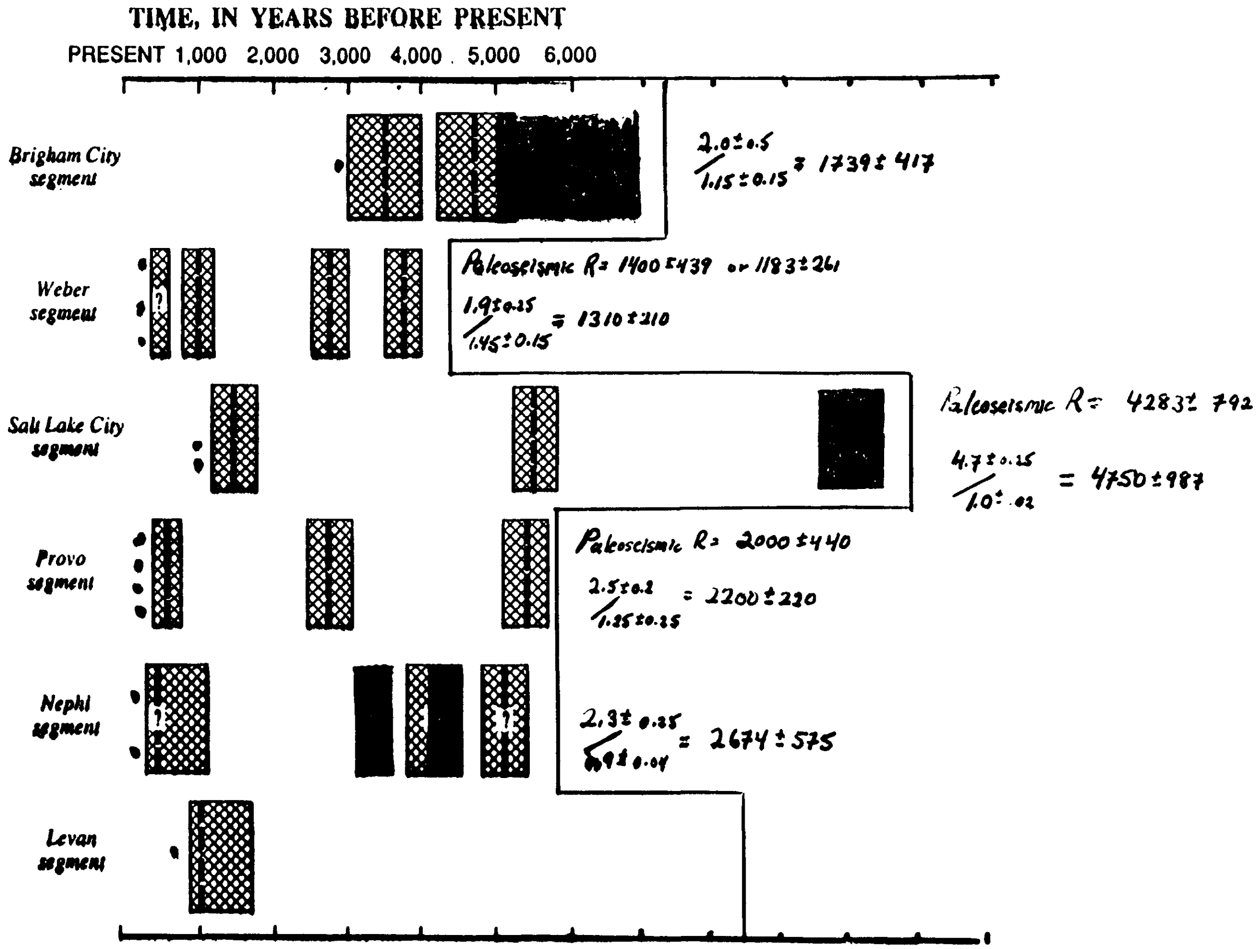

Figure 4: Space-time diagram of prehistoric earthquakes along the Wasatch fault zone, Utah. Boxes represent individual surface rupturing events along the 6 principal fault segments. The width of the individual boxes bracket the range of permissible dates for events, vertical dashed lines are the best estimates of individual event dates, question marks denote those events that are poorly constrained in time (D.P. Schwartz and S.P. Nishenko, written communication., 1991).

S.NISHENKO with the USGS in Golden presented preliminary results of the application of different recurrence and probability models for the Wasatch front. One model, based upon a Poisson or exponential distribution, assumes events are unrelated and uses the minimum information. The second model is conditional and uses details derived from the systematics along the fault, including segmentation, time between events, and time since the last event.

Over the past 6,000 years, a total of 15 to 17 earthquakes with magnitudes ≥ 7 have occurred on the Wasatch. For time intervals of 50 years and 100 years (P_{50} and P_{100}, see Table 1), one obtains 12 to 13 percent and 22 to 25 percent probabilities for a similar event occurring anywhere along the Wasatch fault. The fault appears to have entered an "active" period about 1400 years ago (see Fig. 4) during which 5 to 6 earthquakes have ruptured 5 of the 6 segments. Assuming we are still in this active period, the resulting probabilities are P_{50}=16 to 19 percent and P_{100}=30 to 35 percent. For both time intervals, the probabilities only address the entire fault.

Table 1 -- Poisson probabilities for earthquakes accompanied by surface rupture along the Wasatch fault zone, Utah. (Data from S.P. Nishenko and D.P. Schwartz, written communication, 1991.)

Paleoseismic data	P_{50}	P_{100}
15-17 events in the last 6 ka 353-400 yr repeat time	0.12 - 0.13	0.22 - 0.25
5-6 events in current active cycle (last 1.4 ka) 233-280 yr repeat time	0.16 - 0.19	0.30 - 0.35

Using the model for faulting of the segments presented by D.P. Schwartz, and assuming quasi-periodic recurrence behavior, one can derive estimations of probability for activity on specific segments (Table 2). For example, for the Salt Lake City and Provo segments, the recurrence intervals and the dates of the most recent events yield P_{50} and P_{100} of 1 and 2 percent, respectively, chances for a similar future event.

R.WESSON interjected that the assumption of quasi-periodic activity is crucial to these low probabilities. The assumption that the data representing significant surface offsets represents the total number of earthquakes that could cause significant damage in the region needs to be carefully addressed. He strongly asserted that the data represented a minimum and that a magnitude 6.5 event would not be represented in trenching data. Because it took some amount of clarification among the cognoscente here to agree to this point, Wesson expressed concern that numbers of the sort just presented could easily be misinterpreted and have negative social ramifications. Perhaps there are ways to learn about this class of earthquakes, for instance with paleoliquefaction studies, in order to obtain a better estimate of seismic hazard.

S.NISHENKO thanked Wesson for clarifying that he was addressing characteristic earthquakes and agreed that the level of understanding of events smaller than those under discussion was quite limited.

Table 2 -- Time dependent probabilities for M > 7 earthquakes along various segments of the the Wasatch fault zone, Utah. Recurrence time estimates in brackets are based on either ^{14}C or direct calculations. Probability estimates are for 50 and 100 year time windows (P_{50} and P_{100}, respectively). (Data from S.P. Nishenko and D.P. Schwartz, written communication, 1991.)

Paleoseismic data	P_{50}	P_{100}
Brigham City segment		
Last event 3600 ybp		
Elapsed time twice the observed		
"Direct" Interval (1739 yr)	0.06 - 0.11	0.12 - 0.20
Doublet (3.5 m at 4 ka)	0.04 - 0.05	0.07 - 0.11
Weber segment		
Last event 500 or 1200 ybp (?)		
500 ybp (^{14}C, 1200 yr)	0.04 - 0.12	0.04 - 0.23
500 ybp (Direct, 1310 yr)	<0.01	<0.02
1200 ybp (^{14}C, 1400 yr)	0.05	0.10
1200 ybp (Direct, 1310 yr)	0.07 - 0.11	0.14 - 0.21
Salt Lake City segment		
Last event 1435 ybp		
^{14}C Intervals (4300 yr)	<0.01	<0.01
Direct Interval (4750 yr)	<0.01	<0.01
Provo segment		
Last event 600 ybp		
^{14}C Interval (2323 yr)	<0.01	<0.01
Direct Interval (2000 yr)	<0.01	<0.01
Nephi segment		
Last event 400 ybp		
"Quiet" interval 2-3 times longer than "Active" interval		
(1.0 - 1.4 ka vs. 2.9 - 3.4 ka)		
If currently in another "Active" interval that		
started 400 ybp, use 1200 yr interval	<0.01	<0.02
Direct Interval (2674 yr)	<0.01	<0.01

D.SCHWARTZ pointed out that, for years, workers would not have been surprised by the occurrence of a magnitude 7 earthquake anywhere on the Wasatch fault. These data indicate that it is unlikely that magnitude 7 events will occur on some parts of the fault in the foreseeable future. That doesn't mean that we won't witness damaging earthquakes on other structures, but those must be discussed and qualified in a different manner.

T.HEATON interjected that statements such as the one Schwartz just made (..."it is unlikely that magnitude 7 events will occur on some parts of the fault...") are the strongest that NEPEC makes. NEPEC recently made a similar statement for the north coast segment of the San Andreas fault (..."it is very unlikely that we can have another earthquake"...), and Heaton expressed his concern that NEPEC may repeat such a statement here. He asserted that these statements have very little statistical or physical basis and that, while such events may seem impossible, they do seem to happen.

W.BAKUN stated that he is more troubled by the probabilities for 6.5 magnitude events in the Salt Lake City area, which dwarf the risks associated with the far more infrequent larger events.

S.NISHENKO continued his discussion of the Provo segment which has had 6 to 7 events younger than about 13,500 ^{14}C years, based upon reinterpretation of work done at Hobble Creek in 1979. This would be consistent with a 2,000 year recurrence interval in a quasi-periodic model.

This suggests that perhaps the Provo and Salt Lake City segments are operating in a quasi-periodic manner for earthquakes with magnitudes ≥ 7. If so, this indicates that there is a small chance of a near-term repetition of the events represented in the geologic record along these two segments.

Looking at the segments north and south of these two segments, however, one finds the suggestion of a different type of behavior, which can be addressed using a cluster model in which two, perhaps three, events occur closely spaced in time and are separated from the next such cluster by a long time interval. This seems clearly to be the case in Brigham City segment, and, perhaps, in the Nephi segment. Another possibility is that the uncertainty of the dates of these events does not permit distinguishing between periodic or aperiodic behavior. Nishenko suggested that one could construct a physical model or argument that this normal fault system is pinned at its northern and southern terminations, thus producing different, perhaps more regular, behavior in the center.

The last event on the Weber segment, suggested to have occurred about 500 years ago in one of five trenches, presents an interesting problem. The three older events on the segment yield a 1400-year recurrence interval. If the youngest event did not occur, about 1,200 years have elapsed since the last event. If the youngest event did occur, than a lower expectation would exist for activity on the Weber segment in the near future. This identifies a specific question that could be addressed by a modest amount of new work: was there an event on the Weber 500 years ago? A factor of 2 difference in probability exists between these two scenarios. With an event occurring 500 years ago, P_{50} is about 2 percent and P_{100} is about 4 percent; with the most recent event occurring 1200 years ago, P_{50} increases to 5 to 10 percent and P_{100} to 10 to 21 percent. The range in probabilities reflects a variation in formal uncertainties, (both parametric and intrinsic) in the recurrence estimates.

The absolute numbers are not critical, but the relative numbers are important. For the Salt Lake City, Provo, and, most likely, the Nephi segments, a fairly low probability exists for a magnitude 7 event in the near future. There also may be a low likelihood for the Weber, as well, with the uncertainty about the most recent event significantly effecting the probability.

A fairly high probability exists, however, for the Brigham City segment. Using a quasi-periodic model, a 2-meter average displacement for the two oldest events, and the best estimate of the slip rate, a direct estimate yields a recurrence interval of about 1700 years. If the last event occurred 3 to 4 thousand years ago, we are about a factor of two longer than the calculated return time. This yields a P_{50} of 6 to 11 percent and a P_{100} of 12 to 20 percent with high uncertainty. Application of a cluster model is also permissible, with the two events closely spaced around 4,000 years and totalling about 3.5 m of displacement. At 1 mm per year, we have come through a quiet time during which 4 m of potential slip have accumulated, and we would now have higher expectations for that segment. This yields a P_{50} of 4 to 5 percent and a P_{100} of 7 to 11 percent.

In summary, a blanket Poisson-based probability model for the entire Wasatch fault provides very little information for the behavior of specific segments. Available data allow the use of time-dependent, segment-specific models for larger earthquakes. The various scenarios presented here illustrate how more, and hopefully better, data could improve our perception of the earthquake hazard along the Wasatch fault zone.

L.ALLISON, Utah State Geologist, initiated a discussion of earthquake hazards in Utah from political and policy perspectives by pointing out that the level of understanding about earthquake hazards has increased significantly in the past decade. In particular, cooperation among the USGS, UGS, and the UU, led to the unqualified success of the 5-year National Earthquake Hazards Reduction Program (NEHRP) that was undertaken in the mid-1980's in Utah. A partial list of accomplishments includes improved maps of active faults, assessment of recurrence intervals for movement along the faults, identification of liquefaction hazard zones, and, notably, getting local governments involved in the process through the county geologist program.

This program resulted in the enactment of county ordinances and the production of maps, and Salt Lake County continues to retain a geologist. Clearly, a lot of what we know about the earthquake hazard in Utah today results from the NEHRP program, but the State is still significantly years behind California in terms of understanding of, and preparation for, a major earthquake. The lack of knowledge in a variety of technical areas and the disinclination of society in Utah to deal more aggressively with the issue lead to the conclusion that Utah is not ready to seriously address credible scientific earthquake predictions.

For instance, the likelihood that the next big earthquake may be on the Brigham City segment was presented at a meeting last year. This was received with dismay in the Brigham City area, and the real estate community was quite upset. Incredibly, though, in Salt Lake City, the feeling was one of relief. The level of sophistication needed to comprehend the fact that an event near Brigham City will impact communities up and down the front just does not exist. Thus, we need to take great care in what we say and how we say it.

Although a lot about what has happened along the Wasatch fault in the past 6000 years is known, that structure is one of a series of active faults in the eastern Basin and Range. Allison suggested that perhaps it shouldn't be separated from the whole family of active faults. A map of Quaternary and Holocene faults for the State shows that detailed knowledge is available for only a small proportion of such structures. There appears to be a difference between tectonic patterns for the Quaternary and for the Holocene.

Although the recurrence intervals on the Wasatch fault have been amply described today, at least another dozen faults or so in the ISB are known to have had one or more Holocene surface faulting events. Allsion estimated that some 90 surface-faulting events have occurred in a restricted part of the ISB over the past 15,000 years, yielding an average recurrence interval there of about 170 years. Yet we don't know slip rates, recurrence intervals, or much about the paleoseismology for these faults, and such information is critical for producing earthquake prediction scenarios.

What type of prediction techniques, if any, should we be looking at for Utah, Allison asked. Forecasts are beginning to be made for the Wasatch fault, but he expressed concern about whether it is appropriate to look at this single fault, or whether the whole package of faults should be addressed. Since the forecasting techniques have been developed in a compressional regime, are they appropriately applied in this extensional region, he asked.

The State of Utah has no plans for making earthquake predictions. Identifying and mitigating the hazard in the State has a high priority. We need to understand more about the various tectonic settings, each of which has unique fore- and after-shock patterns. No reliable precursors have been identified. Recurrence intervals are variable, and there appears to be little relationship between faults and earthquakes.

Allison and two others (Lorayne Frank, Director of the Utah Comprehensive Emergency Management Division, and Walter Arabasz, UU Seismograph Stations) have been meeting regularly for a year and a half to coordinate activities and move policy forward. Allison outlined a proposal to obtain authorization for a Utah earthquake advisory board, briefly discussed the possibility of a Utah earthquake prediction group, and asked advice concerning how one might be set up and operated. The need may already exist.

The UGS had to respond to the Browning prediction last Fall, when local schools were closed, parents kept children at home, and many left the area temporarily. The State's response was ad hoc, but relatively successful. Other equally unscientific local predictions, based on lake level, are volunteered on a monthly basis. The UGS is called upon to address these "predictions," but the group is concerned about how it would address a credible forecast or prediction.

In order to partially fulfill their responsibility to educate the populace and leaders about earthquake hazards in Utah, a number of interested experts prepared a report (Arabasz, 1991) and labeled it a "Consensus Document." This document (Appendix E) was presented to NEPEC for consideration.

Finally, Allison opined that there may be a need for a prediction evaluation group in Utah. NEPEC could help by presenting problems and concerns that have been addressed elsewhere, and how they have been dealt with, in addition to providing the broader technical expertise. Understanding how NEPEC works, interacts with the States, and has dealt with problems would be beneficial. How would the State and NEPEC be affiliated? Would there be an exchange of delegates, and could USG call on NEPEC for help if a difficult problem arose?

R.WESSON addressed some of Allison's questions by presenting a brief history of the California Earthquake Prediction Evaluation Council (CEPEC). The Council was formed in the early to mid-1970's, in part to address specific predictions such as that proposed for the Los Angeles region, as an operative agency of the California Office of Emergency Services (OES). As chair of CEPEC, Jim Davis, Chief Geologist for the California Division of Mines and Geology, actually reports to OES for purposes of CEPEC activities and, for the most part, CEPEC responds to issues brought to it by OES.

NEPEC was formed in about 1979, to advise the Director of the USGS, who was given the responsibility to issue earthquake predictions by the National Earthquake Hazards Reduction Act of 1977. The nature of predictions to be issued was not specified, and both narrow and broad interpretations of the mandate have been applied, depending upon the circumstances. Only specific predictions, such as the prediction for Peru, were addressed in the early days of NEPEC.

In the mid-1980s, it was determined that NEPEC could perform another important function, which is to address statements that were beginning to be promulgated concerning future events. Some of these statements were probabilistic and some were not. For instance, one group would assert that there was a 20 percent chance for an earthquake on a certain segment of the San Andreas, and another group would say that a 3 percent probability existed over a certain period of time. Thus, NEPEC became a forum for review, discussion, and consensus evaluation of some longer-term hazard statements.

NEPEC addressed the likelihood of a subduction zone earthquake in the Pacific Northwest. The USGS Director informed the State that there was not yet a complete consensus within the scientific community that such an earthquake was possible, but that the USGS and NEPEC nevertheless considered such an event credible and one that should be taken seriously.

Through the 1980's in California, a number of probabilistic estimates were made about different segments of the San Andreas. NEPEC set up a working group to review all the probabilities and produce a systematic consensus estimate for the fault. This was released in 1988. Some investigators feel that the 1988 document included a forecast that was fulfilled by the 1989 Loma Prieta earthquake. Following that event, and in light of new data, NEPEC convened a working group to review the probabilities for the San Francisco Bay Region and come up with a new consensus document representing the communities best estimate.

It was an early realization that a considerable amount of NEPEC's activities would be centered on California, so the Chairman of CEPEC has always been a member of NEPEC. NEPEC has always been sensitive to the need to confer with CEPEC before going public with a statement.

In sum, two main reasons justify a council such as NEPEC. One is to respond to specific responsible predictions and selected nonscientific predictions. The second mission is to obtain scientific consensus about longer-term hazard, in the 10 to 50 year range. Thus when considering what Utah should do, one needs to think about these issues. If Utah wanted to form a council, NEPEC would be pleased to work out some collaborative mechanisms .

T.HEATON added that California has found it useful to have independent groups from which it might obtain advice, in particular the CEPEC and California Seismic Safety Commission (CSSC), as well as NEPEC. He urged that Utah not name its body the "Utah earthquake prediction evaluation council," because that might actually bait some members of the broader community, but call it something like "Utah earthquake advisory council." One of the important functions of CEPEC is that of providing emergency advice to the Governor in the event of an earthquake sequence. Earthquakes are good times to educate emergency services agencies and public officials. Without an advisory council, such opportunities are lost.

W.BAKUN noted that generally the earthquake advisories associated with moderate events have been positive and have been used both for public education and to get local officials to consider what they might do in the event of a larger earthquake. The two that were issued in the southern San Francisco Bay Region the year or so before the Loma Prieta event were useful in helping city and county officials to perform well in response to that larger earthquake.

J.DIETERICH addressed Allison's view that it might be premature to undertake a probabilistic assessment in Utah. Although the point might be a valid one, the predictions prepared in California have ended up having a dual purpose. One is a societal or political purpose and the other is scientific. The reports have all had strong language that they are status reports of current thinking and that the consensus will change. While they are obsolete by the time they come out, the prediction statements really help various parts of the community focus and take action. Preparation of the reports focuses future scientific activities.

JUNE 11, 1991
Evening Session

R.SMITH, L.ALLISON, and **W.ARABASZ** led a panel discussion which was started by addressing three items:

1. State coordination with NEPEC;
2. Needs of probabalistic analysis for Wasatch; and
3. Recognition of need for work on earthquake forecasting problems on Wasatch "normal' fault mechanisms.

R.SMITH briefly discussed earthquake research and policy in Utah and described the need for some organization to lead information gathering and public dissemination. He agreed that establishment of a "prediction council" in the State at this time would be premature.

W.ARABASZ indicated that the process of public education was a long-term process and that the small number of working professionals in the UGS and UU somewhat limited the group's respect in the eyes of local politicians and the leaders of powerful state agencies. Having a national group support the findings of the local experts would be very helpful.

T.HEATON opined that he doubted that any policy apparatus or group that would respond to a sequence of normal faulting events existed in the State. This would be a natural thing to consider. Once a damaging earthquake occurs in the state, the public will insist that the governmental leaders do something to address future events.

J.DAVIES pointed out that one of the useful features about the earthquake community in California has been the consensus concerning the earthquake situation, whether an earthquake "forecast" or "prediction." A reasonable prediction scenario that has some national authority behind it fosters mitigation efforts and lends credibility to local workers. Perhaps working toward some sort of document that might have the NEPEC "stamp of approval" might be a useful endeavor. Some discussion ensued whether NEPEC should provide such a stamp.

W.ARABASZ took this as opportunity to further describe the "consensus" document presented to NEPEC during the meeting. At the suggestion of Walter Hays (USGS, Reston), prior to a workshop on the Wasatch front, several working groups formed to develop and articulate a consensus view about aspects of the earthquake hazard in Utah. To date, the document has had limited distribution.

R.WESSON focused the discussion somewhat. The NEPEC process represents a small part of the earthquake milieu. The focus of the process is not isolated from politics, but the primary focus is on the scientific aspects. Any response can be tuned to have the correct balance of external and internal contributions in order to attain the desired scientific and political outcome. He asked whether this was the appropriate time, scientifically in terms of our understanding of the problem, to bring together a summary, or should we wait?

R.SMITH pointed out that because of the NEHERP effort, groups with earthquake hazards expertise and interest exist in the State and the USGS, and that, regardless of the uncertainties that remain, this may be an opportune time to address the issue in a summary fashion. If we wait too long, we risk losing interest and momentum.

A.JOHNSTON pointed that since Schwartz and Nishenko are preparing a paper on the Wasatch fault zone with a time dependent focus, perhaps a similar analysis could be done for other parts of the state.

K.SHEDLOCK queried whether enough information existed for such work elsewhere in the State. Further, she asked whether a distillation document or the consensus document would be the best item for NEPEC to endorse.

T.HEATON wondered about the efficacy of a document stating that NEPEC has reviewed the program and concurs that a real earthquake threat exists. He asked if such a statement wouldn't urge the process along?

T.McEVILLY proposed that an alternative would be for NEPEC to report to the Director the culmination of an intense and successful program that has lead to estimates of the hazard in Utah and to propose that now might be the time to suggest folding the resulting information into State policy activities.

A consensus emerged at this point that a plateau has been reached and that enough information existed to prompt action on the part of the State, and that NEPEC somehow needed to express the situation and that level of understanding.

J.DAVIES opined that while it would be appropriate for NEPEC to review the consensus document, assess it, and communicate the consensus to the Director, it would not be appropriate for the Council to tell Utah what its response to the document might be.

R.WESSON stated that some sort of probabilistic estimates will be forthcoming soon, but that unresolved scientific issues remain. In particular, a serious disconnect exists between the geologic element, which appears to be self-consistent and complete, and the seismological element. In a year or so, a probabilistic forecast that would include both elements could be produced.

It would seem to be within the capability of the community to address the issue over the next year and then to stimulate the continuing process. It might not be enough just to review the consensus, or to accept a paper that might be written on the Wasatch fault, but the Council should address all the data and come up with a complete synthesis and description of the hazard.

J.DIETERICH suggested that a communication expressing the end of the federal program in Utah allows a synthesis of the information produced by the program and provides a preliminary evaluation of the earthquake hazard for the State.

W.ARABASZ stated that the process would be stimulated by recognition of the problem and assimilation of the information into policy activities. NEPEC is in a position to affect policy. He would like to see a communication encouraging the next step for UGS, UU, and the USGS.

K.SHEDLOCK pointed out that if this were the case, the consensus document is a starting document for the communication. This would mean that workers from Utah need to participate in the working group, which would be the next step.

Consensus evolved to one that involved a NEPEC evaluation of the "consensus document" and supplemental information that may include some time-dependent models. The resulting conclusions about the document would form the basis of a letter written by McEvilly to the Director.

JUNE 12, 1991
Morning Session

J.LANGBEIN presented a review of the B-level alert at the Parkfield Prediction Experiment in March, 1991. He summarized by stating that the alert most likely resulted from a combination of rain-induced events recorded by two creepmeters and perhaps a typographical error or oversight in the protocol for initiating alerts.

R.WESSON briefly interjected that he found the B-level alert to have been entirely appropriate under the circumstances and given the rules of operation. Several panel members concurred, stating that the caveats announced with the alert seemed appropriate.

J.LANGBEIN continued by presenting illustrations of creep measurements and rainfall. He presented the rules (Bakun, W.H., and others, 1987) under which the alert was made and discussed some of the uncertainties, including the possibility that a limiting phrase referring to "confirming signals" was omitted from the description for status B(2) that was included with status B(1). He proposed changes that would tighten the rules including some concerning the term "tectonic origin." These suggestions were incorporated in a June 10, 1991 memo (Appendix F) distributed to the Council.

Much discussion ensued, and opinions varied depending upon the individuals experience and expertise. Concern was expressed that an alert not be made on the basis of creep alone. Others stressed the rules must be clear. One or two stated that they would be comfortable with a specific rule for events occurring during periods of high rainfall. The Council encouraged Langbein to more formally propose a revised protocol reflecting the day's discussion to account for the rainfall problem, as well as other changes, and to circulate the proposals to the Council by memo before the next meeting.

J.LANGBEIN agreed. He next expressed concern about low-frequency instrumental response, especially with regards to the instrument thresholds and noise. The fact that we see a signal from an instrument does not necessarily mean that the probability of an earthquake has been increased because we have no statistics relating the occurrence of deformation with an ensuing earthquake.

Langbein has been considering readdressing the issue of thresholds for all instruments using a "match-filter" technique, which requires some understanding of the power density spectra of the instruments. (See discussion, Appendix F.) This would allow statements about the confidence that a given signal exceeds the background noise of that instrument.

Some discussion ensued. The Council concluded that while the "blue-book" is a known quantity, a match-filter system may be more rational, and that it is appropriate to look at the possibilities. Langbein was asked to document his analysis of instrument thresholds, to determine what might have been done differently using the match-filter technique during past C and D alert levels, and to propose a protocol that might improve the blue book for Parkfield and serve as a better model for a similar document for the San Francisco Bay Region. The Council agreed to readdress Parkfield at the next meeting.

J.DIETERICH discussed probabilistic earthquake forecasting techniques used recently in California and proposed for use elsewhere. Characteristic earthquakes, segmentation, and recurrence models are the building blocks for these time-dependent models. Each of these elements has a number of unresolved issues, and many open research questions remain available for investigation. Dieterich expressed concern about the enigma that the techniques are more difficult to apply in regions with lots of information than in regions with sparse information.

A recent letter, written to Wesson by A. Cornell, expressing concern about the intrinsic coefficient of variation, how well the coefficient is established, and its impact on uncertainty was discussed by Dieterich. Cornell proposed a small group of practitioners to determine points of agreement and disagreement and produce a "white paper." A second initiative proposed by A. Lindh and S. Nishenko would involve a "red book" conference on earthquake probabilities. By including regional issues, such a conference would go beyond methodology.

R.WESSON, in reply to a comment made by Heaton, stated that using logic tree models to deal with uncertainty addresses a number of important issues and doesn't obfuscate the messiness. The issue of most concern to Wesson is the characteristic earthquake assumption. NEPEC likely will be pressed to produce more probabilistic estimates and in doing so needs to use the most frontier methods to reflect a consensus of the best thinking. Now is the time to bring the probabalistic gurus together to address the issues and communicate their areas of agreement and disagreement. This should be a small group focussing on the technique.

J.DIETERICH suggested that a small group meet in advance of a "red book" conference and present a report at the outset of the conference.

T.McEVILLY pointed out that the vehicle for such a study exists as a spin-off of the 1990 report. Cornell has offered a study he is involved in as the vehicle, and he wants physics and seismology to help drive the report. The working group should involve the chief practitioners.

W.PRESCOTT agreed that the Council must use the best criteria and methodology available to maintain credible results, but he was concerned that as NEPEC addresses issues in other areas, the Pacific Northwest or Wasatch front, for instance, the different environments might require a more locally appropriate methodology. How generic is the methodology?

J.DIETERICH concurred that it is important to recognize the inherent differences in our level of knowledge in different regions. The idea is to convene the group to discover both areas of consensus and areas with problem.

K.SHEDLOCK suggested that this was particularly important in light of the mild discomfort resulting from necessary hasty use of parameter values in one report recently sanctioned by NEPEC. All such issues should be readdressed in order to define a consensus and to make them as "clean" as possible, particularly in areas where significant amounts of data exist, before the Council addresses similar issues in areas with a less robust data base. The proposal for a small methodological working group to distill their thoughts is a very good idea, and the group should include chief critics.

R.WESSON stated that the small group of people addressing the methodology issue might meet within the next few months. The second, more broadly based, workshop should be considered independently. It's lifespan would be 9 months to a year from organization to final report. A carefully written charge for the 2 to 5 day meeting would focus the participants attention on three items: characteristic earthquake, recurrence, and segmentation. The group should also report on unresolved issues that need additional work.

Jim Dieterich and John Davies were selected as chairmen of the workshop. They were asked to draft and circulate a charge for the workshop and to also circulate a list of possible participants.

K.SHEDLOCK presented the group with a draft of USGS Circular 1067 (Shedlock and Weaver, 1991), which summarizes the work that has been done thus far in the Pacific Northwest. The document is one of the results of a workshop attended by 14 invited participants that was convened to formulate a regional earth science plan for earthquake hazards reduction.

The document discusses the possibility of an 8 to 9 magnitude earthquakes, but it concludes that the probability of such an event would be difficult to establish. For instance, some of the studies of terraces and uplifted marshes may be indicating segmentation of the plate boundary, perhaps reducing the possibility of such a great event. Shedlock stated that one or all of the segments might produce a magnitude 8 or greater event, but pointed out that workers involved in the research, as well as reviewers of the document, have yet to reach a consensus that would allow a more robust statement.

In any event, such an earthquake may not be the most hazardous -- crustal urban earthquakes near urban centers may prove to be more hazardous. The hazards from each of the three possible types of events, intraplate, crustal, and subduction earthquakes, still need further definition, making consideration of deterministic probabilities premature.

A.JOHNSTON pointed out that the question has shifted from "is the interface seismic and capable of producing a big earthquake" to "does segmentation of the interface preclude a magnitude 8 earthquake." This represents a big advance.

R.WESSON stated that the Council probably was not ready to debate the issue, but, instead, should decide upon the process that it will use to make a summary NEPEC statement for the Pacific Northwest by focusing on what is agreed upon and then moving on to the areas of disagreement.

Much discussion ensued. Members concluded that most workers would accept the possibility of an 8 magnitude, but the possibility of events larger than 8 lacks consensus. The Council agreed that K. Shedlock, C. Weaver, and H. Kanamori would prepare a short draft NEPEC statement (summarizing issues for which a consensus exists and issues that remain unsolved) of the situation in the region and circulate it in advance of the next NEPEC meeting at which the most recent finding would be presented.

T.HEATON opened a discussion on earthquake probabilities for southern California by presenting the question whether or not the blue book probabilities volume for the region that was produced in 1988 should be updated for southern California as was done for the San Francisco Bay region following the Loma Prieta earthquake. A number of issues bear on the question:

- -The original report covered the San Andreas and San Jacinto faults, but not the Elsinore, Newport-Inglewood, nor ,in particular, blind thrusts.
- -Methodologies have changed.
- -Shaking probabilities would be effected by changes in faults considered and in methodologies.

The proper organization to discuss these questions was the Southern California Earthquake Center (SCEC), Heaton asserted, and he saw a conflict between a NEPEC working group and SCEC on the issue. He pointed out that one of SCEC's main objectives was the production of a "master model" of shaking hazard, and the likely distribution of earthquakes is an important component of that model, as well as the problem of other faults and hidden thrust faults. He felt that it would be politically difficult for both SCEC and NEPEC to separately address the issue. Perhaps it would make sense for NEPEC to communicate with SCEC, stating that NEPEC is anxious to have the results of that study and to try to schedule a report on the progress in developing that model in a year or so.

Various SCEC working groups have been formed, Heaton continued, and one has to do with the master model, which will evolve with time. The model likely will start with a modification of Steve Wesnousky's shaking hazard map. Underlying such a study is the suite of possible earthquakes. Another working group is providing information for this question. These may result in a model of the activity rating of faults, a model of how the seismicity can be distributed from that activity, a model of how the waves travel through southern California, and a model of how these parameters can be unified into the "master" shaking model.

T.McEVILLY disclosed that in discussions with Thomas Henyey, SCEC Executive Director, and Keiiti Aki, SCEC Scientific Director, the two expressed the desire to operate within a NEPEC charged working group rather than some vehicle operating out of the center.

H.KANAMORI stated that, in his opinion, we were not talking about joint work on the the master model, which is a scientific product, but about jointly addressing the probability issue, while also addressing the societal impact.

T.HEATON expressed concern about a bluring of boundaries between SCEC, NEPEC, and the USGS, as well as about the timing of various products that may originate from SCEC and the USGS.

R.WESSON asserted that the master model, the Center's research objective, arguably will not reach a plateau for 5 years, if ever. NEPEC's responsibility is to do a short-term, one-year, study to update where we stand. How do probabilities differ between 1988 and now? Enough new information exists to require such an update, and it seems that a working group involving SCEC and NEPEC could update the probabilities without obviating the need for a completed master model.

Discussion continued in this vein. **W.PRESCOTT** summarized the concerns of several members that NEPEC not delegate it's obligation to update the probabilities in southern California to another organization over which the Council has no control. After a resummary of the discussion between McEvilly and SCEC's two directors, **T.McEVILLY** stated that it seemed to him that the confusion surrounding establishing a joint working group was a turf problem. When it was suggested that the results of the working be made available next summer, **T.HEATON** asserted that it would be premature to update any of the three questions presented at the outset.

D.SCHWARTZ offered the opinion that adding the Elsinore or Newport-Inglewood faults would slightly increase the probability, just as adding the Rogers Creek fault did in the San Francisco Bay

Region. The real changes in the probabilities will occur due to the addition of a methodology quantifying recurrence intervals for the blind thrusts. This is the critical issue, and a group should be established just to address and develop this issue and methodology.

J.DIETERICH pointed out that probability reports are always premature. They are written in response to the need to provide some sense of what we know and transfer that information and numbers to the community at large.

The Council agreed that the Chairman and Vice Chairman should communicate with SCEC's science and executive directors to develop the charge, composition, and timing of a joint working group, which might be cochaired by workers from SCEC and NEPEC. Tom Heaton was designated NEPEC's representative for purposes of the working group.

R.WESSON next reviewed the Evernden ground motion calculations for the San Francisco Bay Area as the first issue under old business. When the working group was nearing completion of the update of probabilities for the San Francisco Bay Region, the idea arose that the earthquakes should be combined with our understanding about how the resulting ground motion would be distributed in the region. Jack Evernden had a program which could be used to produce such ground motion maps. At that time, Jim Davis expressed the desire to have the California Division of Mines and Geology involved in the process.

T.HEATON explained that the "Brown bill" in California mandated that the State delineate special study zones for the hazards related to liquefaction, landslide potential, and ground shaking by this Fall and that the State may be concerned by an apparent overlap in responsibilities.

W.BAKUN stated that at the last NEPEC meeting Members agreed that it would be a good idea to get this type of product out, but that all felt that the maps must be understood and defensible. The Council started the process by asking Evernden to produce the maps. The maps have been produced, but the State, which expressed concern about some of the parameters, has not yet reviewed them, and, as a matter of fact, they have yet to be reviewed by the Survey.

R.WESSON outlined several options, and, as an expedient, the Council agreed that Bakun would draft a letter for McEvilly to send to Davis suggesting that NEPEC feels the need to resolve the issue and that excludes the option to spend more time studying the issue. Without Davis' objection, the group would like to proceed by getting the maps out in open-file form. This would give the State the option to get involved in the process, but also set limits so that the author can get his ideas out for review.

A brief analysis of NEPEC activities with regards to the so-called "Browning prediction" brought comments from some Members that NEPEC should have acted sooner, and it was suggested that the Council mistakenly thought that the Central United States Earthquake Council would initiate an analysis of its own.

G.JOHNSON, of FEMA's Earthquake Hazards Program, suggested that a more timely review of the prediction might have put it to a deserved rest, but that the prediction got so much early media visibility that, without hearing otherwise, Browning gained credibility.

On other issues, **G.JOHNSON** expressed his pleasure for being able to attend the NEPEC meeting. His observations of the deliberations made it clearer then ever that NEPEC was getting more into the public policy arena. He noted that at a recent FEMA workshop, working groups of States (32 States now participate in FEMA's NEHRP program) uniformly presented the need for credible hazard and risk information. He concluded by emphasizing the importance of an intermediate publication on probability in southern California.

The Council closed the meeting at Alta by discussing agenda items for the next NEPEC meeting. Members already had decided to address the Pacific Northwest and Parkfield.

Since Jack Healy has continued to work with Keilis-Borok and co-workers on the TIP model, it was decided that it would be appropriate for him (Healy), Jean B. Minster, Mark V. Matthews, or Stewart W. Smith to present an update, totalling an hour. What have American scientists learned about the technique and how has it been applied?

Likewise, the Council expressed the desire to be updated on the electromagnetic, VAN, prediction method being used in Greece, but decided to wait on such a briefing.

It was agreed that the meeting would be held in Oregon on October 28 and 29, 1991.

J.DAVIES proposed that a working group be set up to revisit the status of gaps in Alaska. **T.McEVILLY** suggested that Davies prepare a proposal for NEPEC and that he circulate it to Council Members before the next meeting. The proposal would include a proposed charge and suggested panel members.

T.McEVILLY informally asked Members to remain on NEPEC unless overriding circumstances necessitated a resignation.

Meeting was adjourned at 12:22 pm.

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(p. 35 follows)

APPENDICES

Appendix A April 29, 1991, letter from Dallas Peck to NEPEC presenting the "charge" for the Alta meeting.

Appendix B Agenda for the June 11 and 12, 1991, meeting of NEPEC at Alta, Utah.

Appendix C Document (Smith and Arabasz, 1991) provided NEPEC by Smith, summarizing seismicity of the Intermountain seismic belt.

Appendix D Document (Machette and others, 1991) presented to NEPEC by Machette, summarizing what is known about the Wasatch fault zone.

Appendix E Document (Arabasz, 1991) presented to NEPEC by Allison outlining a consensus view of activities that would reduce losses due to earthquakes.

Appendix F Memo, dated June 10, 1991, presented to NEPEC by Langbein outlining a proposal to revise the Parkfield earthquake prediction scenario.

Appendix A

April 29, 1991, letter from Dallas Peck
to NEPEC presenting the "charge" for the Alta meeting.

United States Department of the Interior

GEOLOGICAL SURVEY
RESTON, VA 22092

In Reply Refer To:
WGS-Mail Stop 905

APR 29 1991

Dr. Thomas V. McEvilly
Department of Geology & Geophysics
University of California
Berkeley, California 94720

Tom
Dear Dr. ~~McEvilly~~:

As you prepare for the next meeting of the National Earthquake Prediction Evaluation Council (NEPEC) scheduled for June 11 and 12, 1991, there are three issues on which we would particularly like the advice of the Council over the next several months to 1 year. We would appreciate your raising these matters with your colleagues and undertaking whatever kind of processes seems appropriate within the NEPEC framework (i.e., workshops, working groups, etc.) to address them. These issues are:

- Methodology for probabilistic forecasts. The probabilistic forecasts, such as the recent update for the San Francisco Bay Area, seem to be at the cutting edge of both geologic and geophysical interpretation, as well as the application of probabilistic techniques. We would appreciate the advice of NEPEC about what methodological issues should be addressed as this field progresses. In what areas is there consensus among practitioners and in what areas is there disagreement? Given these concerns, what guidelines should be given to future work groups assigned the task of revising probabilistic forecasts?

- Earthquake hazard in the Pacific Northwest, especially that associated with the Cascadia subduction zone. Several years ago, NEPEC reviewed the then emerging evidence for past great earthquakes along the Cascadia subduction zone. At that time, NEPEC advised that this threat should be taken seriously, but that a consensus on the significance of the new information was not complete. Since that time, considerable additional studies and analyses have been carried out. Therefore, we would like to ask NEPEC to undertake an analysis of the current understanding of the earthquake hazard in the Pacific Northwest, including an assessment of the current consensus view on the potential for future great earthquakes along the Cascadia subduction zone.

- <u>Revision of earthquake probabilities for southern California</u>. Since the preparation of the first NEPEC probability report for California issued in 1988, significant new information has emerged about the importance of various types of faulting and about past earthquakes in the region. We would like to ask NEPEC to determine whether this new information is sufficient to warrant revision of the probabilistic forecasts for that region, and, if so, to lead an appropriate process to revise those estimates.

If you have any questions about this request that cannot be answered by Rob Wesson, Chief, Office of Earthquakes, Volcanoes, and Engineering, please do not hesitate to call us. Of course, if there are other issues that NEPEC feels should be raised, please let us know.

We very much appreciate the work you and the other members of NEPEC have done. Your efforts on behalf of the National Earthquake Hazards Reduction Program and the USGS have been extremely valuable. Thank you very much for your help.

Sincerely yours,

/s/ Dal

Director

Appendix B

Agenda for the June 11 and 12, 1991, meeting of NEPEC at Alta, Utah.

National Earthquake Prediction Evaluation Council

Alta Lodge
Salt Lake City, Utah
June 11-12, 1991

Tentative Agenda

June 11

8:30 a.m.	Introductory business and discussion	Tom McEvilly, Chairman Rob Wesson, Vice Chairman
9:00 a.m.	Seismotectonics, seismicity and paleoseismicity of Intermountain Seismic Belt (ISB)	Bob Smith University of Utah
9:30 a.m.	Seismic hazard of the Wasatch Front	Walter Arabasz University of Utah
10:00 a.m.	Geodetic measurements - Intermountain Seismic Belt	Jim Savage, USGS
10:30 a.m.	Geometry and state of stress of Wasatch Front	Ron Bruhn Utah Geological & Mineral Survey
11:00 a.m.	A summary of Quaternary geologic studies, Wasatch	Mike Machette, USGS Bill Lund, Utah Geological & Mineral Survey
11:30 a.m.	Paleoseismic evidence for segmentation of normal faults	Dave Schwartz, USGS
12:00	Lunch	
1:00 p.m.	Probabilistic estimates for the Wasatch fault	Stu Nishenko, USGS
1:30 p.m.	Nucleation processes on normal faults	Chris Scholz, Lamont
2:00 p. m.	Earthquake prediction in Utah: A State Perspective	Lee Allison State Geologist, Utah
2:30 - 5:00 p.m.	Discussion: The Intermountain Seismic Belt (this could include subjects like segmentation, recurrence models, what are the expected precursors to large normal fault events, planar vs. listric, future research/monitoring needs, etc.	

June 12

8:30 a.m. – 12:00 p.m.

1) Revision of the Parkfield alert criteria (John Langbein)

2) Methodology for probabilistic earthquake forecasting: Is it time for a review?

3) Status of the earthquake hazard research in the Pacific Northwest: Is it time for some probabilities?

4) A reevaluation of probabilities for southern California: Is it time for a new working group?

Appendix C

Document (Smith and Arabasz, 1991) provided NEPEC by Smith, summarizing seismicity of the Intermountain seismic belt.

The Geology of North America
Decade Map Volume 1
1991

Chapter 11

Seismicity of the Intermountain Seismic Belt

Robert B. Smith and Walter J. Arabasz
Department of Geology and Geophysics, University of Utah, Salt Lake City, Utah 84112

INTRODUCTION

In this chapter we present an overview of the Intermountain seismic belt (ISB), a first-order feature of the Seismicity Map of North America (Engdahl and Rinehart, 1988). The ISB is a prominent northerly-trending zone of mostly shallow (<20 km) earthquakes, about 100 to 200 km wide, that extends in a curvilinear, branching pattern at least 1500 km from southern Nevada and northern Arizona to northwestern Montana (Fig. 1). Our study area, defined by the bounds of Figure 1, covers a sizable part of the western United States encompassing the ISB and is informally referred to herein as the Intermountain region.

Contemporary deformation in the ISB is dominated by intraplate extension. Forty-nine moderate to large earthquakes ($5.5 \leqslant M_S \leqslant 7.5$) since 1900 and spectacular late Quaternary faulting with a predominance of normal to oblique-normal slip make the Intermountain region a classic study area for intraplate extensional tectonics. Information from the Intermountain region, relating for example to paleoseismology (Schwartz, 1987), seismotectonic framework (Smith and others, 1989), contemporary deformation from geodetic measurements and seismic moments of earthquakes (Savage and others, 1985; Eddington and others, 1987), and strong ground motion in normal-faulting earthquakes (Westaway and Smith, 1989a) has added significantly to understanding extensional seismotectonics worldwide. Particularly valuable contributions have come from field and seismological observations of two large normal-faulting earthquakes in the Intermountain region—the 1959 Hebgen Lake, Montana, earthquake (M_S = 7.5) and the 1983 Borah Peak, Idaho, earthquake (M_S = 7.3)—both described herein. Our basic intent in this chapter is to provide an interpretive guide to the seismicity of the ISB. We also summarize and discuss observations from the Intermountain region that are relevant to general aspects of extensional intraplate tectonics.

The coherence of the ISB as a regional earthquake belt became apparent with evolving compilations of seismicity (Heck, 1938; Woolard, 1958; Ryall and others, 1966). The earthquake belt was well defined in Barazangi and Dorman's (1969) global seismicity map of shallow earthquakes and was first called the "Intermountain Seismic Belt" in joint abstracts by Sbar and Barazangi (1970) and Smith and Sbar (1970). Follow-up papers by Sbar and others (1972), and especially one by Smith and Sbar (1974), gave modern seismotectonic overviews (see also Smith, 1978; Arabasz and Smith, 1981; and Stickney and Bartholomew, 1987).

The ISB roughly follows the eastern margin of a broad region of late Cenozoic crustal extension in western North America. This seismically active boundary with more stable continental interior to the east has been interpreted as a subplate boundary (Smith and Sbar, 1974; Smith, 1978). It is well known that, on a regional scale, the ISB coincides with a persistent deformational belt in western North America that has been recurrently active since late Precambrian time (Levy and Christie-Blick, 1989; Anderson, 1989) and which is now characterized by pronounced lateral heterogeneities in crust-mantle structure across the ISB (e.g., Smith and others, 1989). Contemporary deformation in the region marks a continuation of late Cenozoic extension and volcanism (in the Yellowstone–Snake River Plain volcanic system and in southern Utah), whose various modern stages began roughly 10 to 15 m.y. ago (Anderson, 1989).

Regional-scale earthquake pattern

There is a general north-south regional continuity to the ISB (see the Seismicity Map of North America, Engdahl and Rinehart, 1988), but we can distinguish at least three parts—referred to herein as the southern, central, and northern ISB (Fig. 1)—for convenient reference. These subdivisions of the ISB may be arguable, but we believe distinctive features of the central ISB, as described in this paper, differentiate it from the ISB to the north and south. Referring to Figures 1 and 2 (see also Fig. 3 for additional features and place names), the southern ISB (36° to 42¼°N) coincides with a tectonic transition zone between the Basin and Range province on the west and the Colorado Plateau–Middle Rocky Mountain provinces on the east. In southwestern Utah at about 38°N there is a southward bifurcation of the ISB. A

Smith, R. B., and Arabasz, W. J., 1991, Seismicity of the Intermountain Seismic Belt, *in* Slemmons, D. B., Engdahl, E. R., Zoback, M. D., and Blackwell, D. D., eds., Neotectonics of North America: Boulder, Colorado, Geological Society of America, Decade Map Volume 1.

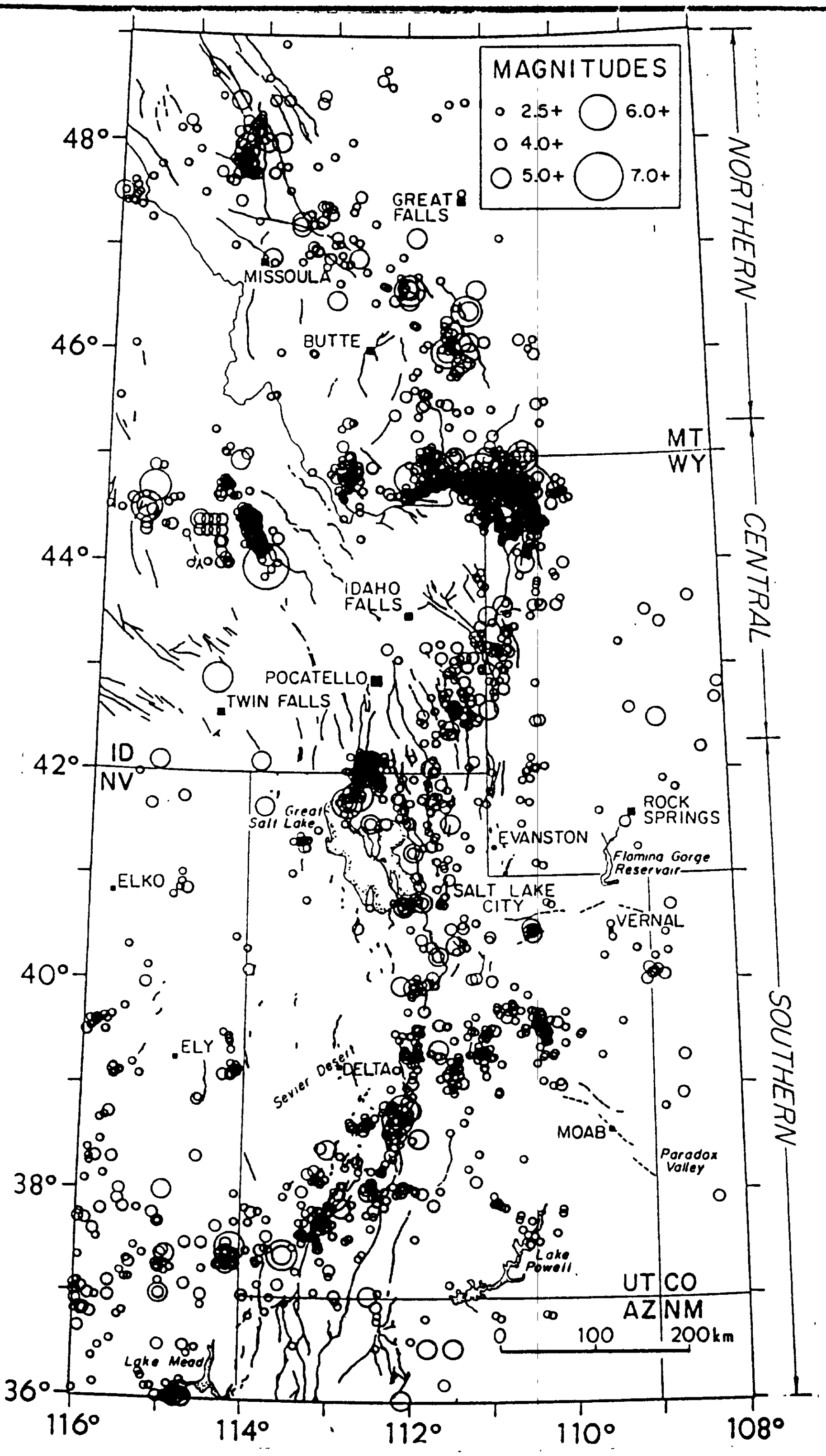

Figure 1. Earthquakes in the Intermountain region, 1900–1985, outlining the Intermountain seismic belt (ISB), together with selected Cenozoic faults identified in Figure 3. Earthquake data are from the compilation of Engdahl and Rinehart (1988; this volume). Northern, central, and southern parts of the ISB are delimited for reference.

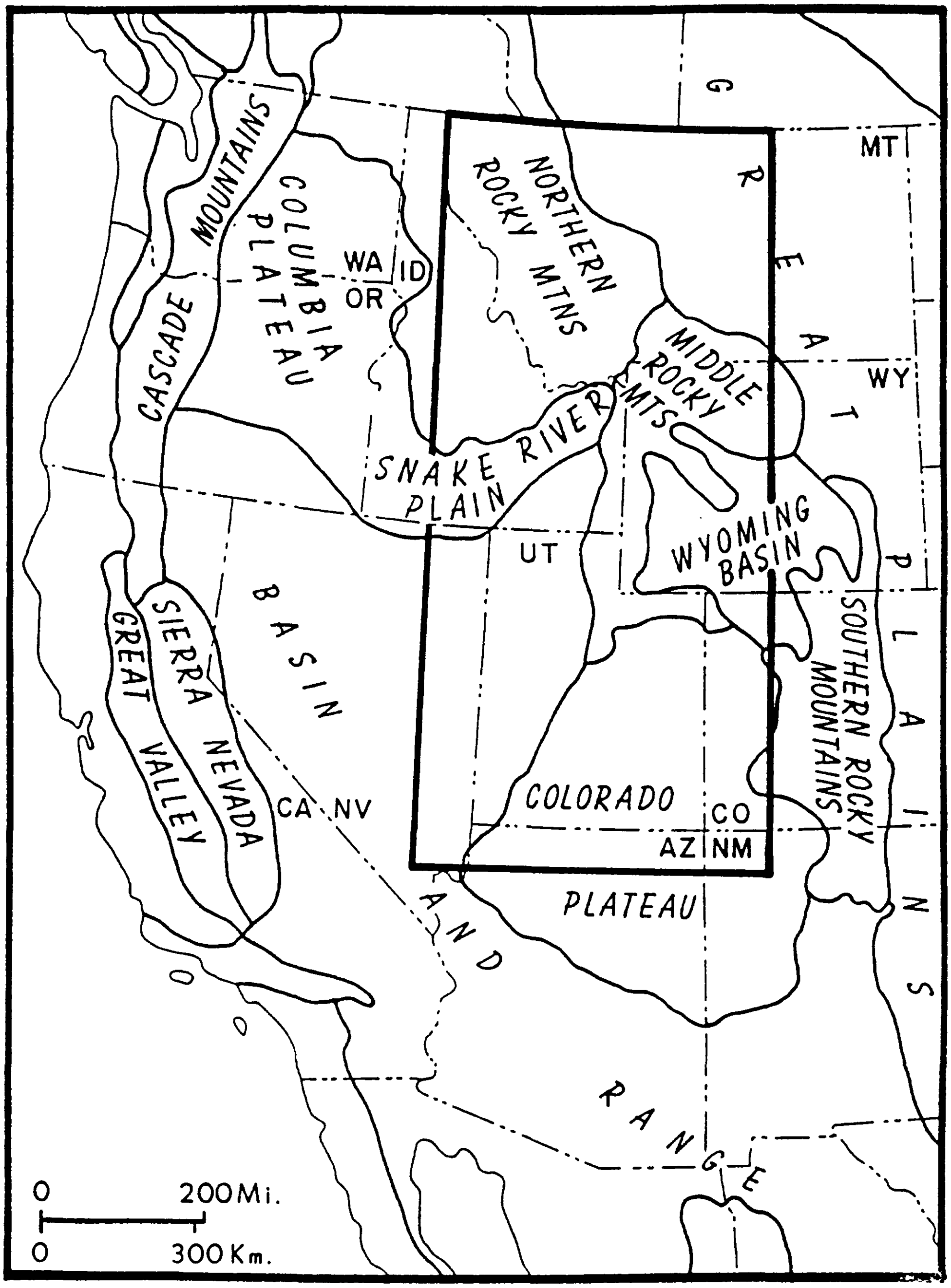

Figure 2. Map of the western United States showing physiographic provinces and location of study area (bold outline) shown in Figures 1, 3, 6, 8, 9, and 20.

distinct belt of seismicity continues southwestward some 200 km across southern Nevada, partly including induced earthquakes related to underground nuclear explosions at the Nevada Test Site in southern Nevada (Rogers and others, this volume); this belt transects the local north-south tectonic grain and coincides with the midpoint of a steep regional gravity gradient (Eaton and others, 1978) between the northern and southern sections of the Basin and Range province. The other part of the bifurcation is a weaker zone of scattered earthquakes that extends southward into central Arizona through a broad belt of Quaternary faulting (see Kruger-Kneupfer and others, 1985, Fig. 3).

In southwestern Utah (Fig. 1), earthquakes of the southern ISB follow a northeasterly structural trend to about 39°N, where both structure and the earthquake belt change to a northerly trend. Clustered earthquakes defining an inverted U-shaped pattern of epicenters in east-central Utah between 39° and 40°N are

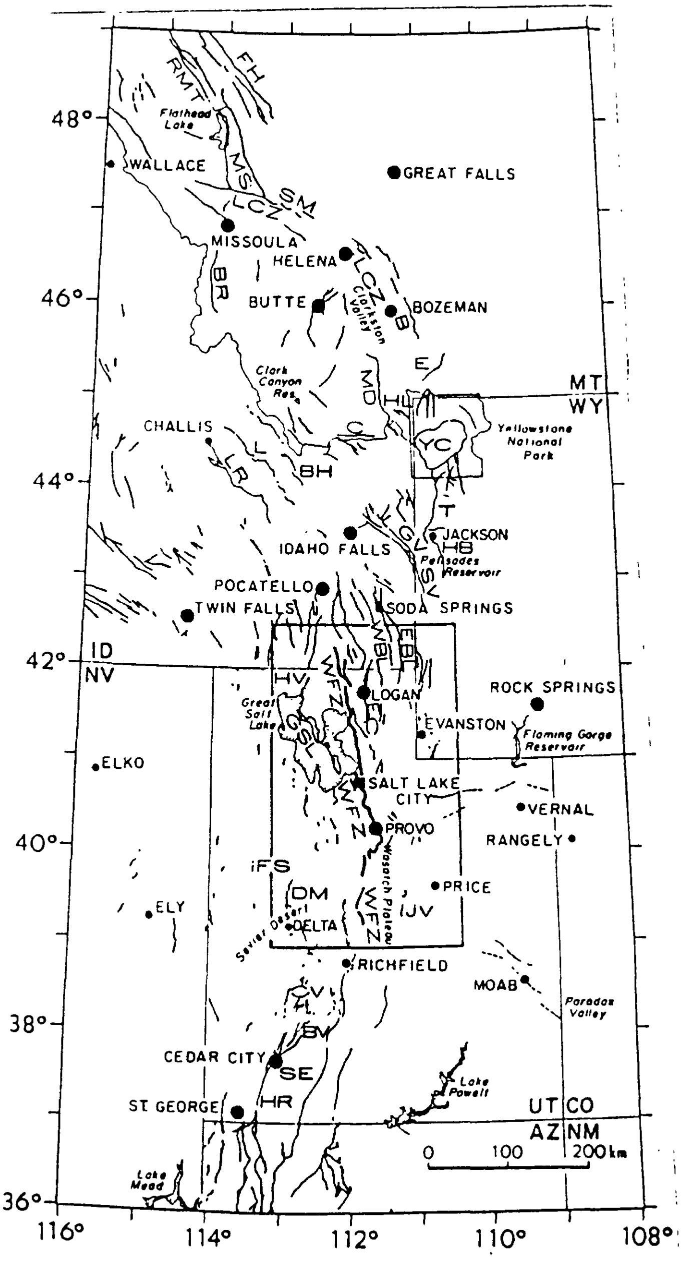

Figure 3. Location map of place names and selected late Cenozoic normal faults of the Intermountain region. Rectangle outlines Wasatch Front area shown in Figure 12, wherein other faults are identified. Abbreviations of faults are as follows: B = Bridger; BH = Beaverhead; BR = Bitterroot; BV = Beaver; C = Centennial; CV = Cove Fort; DM = Drum Mountains; E = Emigrant; EBL = East Bear Lake; EC = East Cache; FH = Flathead; FS = Fish Springs; GSL = Great Salt Lake; GV = Grand Valley; HB = Hoback; HL = Hebgen Lake; HR = Hurricane; JV = Joes Valley; L = Lemhi; LR = Lost River; MD = Madison; MS = Mission; SM = St. Mary's; SE = Sevier; SV = Star Valley; T = Teton; WBL = West Bear Lake. Other labeled tectonic features are: LCZ = Lewis and Clark Zone (see Fig. 18); RMT = Rocky Mountain trench; YC = Yellowstone Caldera.

mining-related (described below). Northward in Utah the ISB centers on the 380-km-long Wasatch fault, the preeminent normal fault zone of the eastern Basin and Range province, along which young mountain blocks have been uplifted to form a major west-facing physiographic scarp, called the Wasatch Front, with up to 2,300 m of relief.

The central ISB (42¼° to 45¼°N) follows, in part, the Basin and Range–Middle Rocky Mountain transition, but is complicated by having a westerly-trending branch; the result is an arcuate pattern that appears to "wrap around" the late Tertiary volcanic province of the eastern Snake River Plain (SRP; see Fig. 2). North of the Utah-Idaho border the ISB takes on a marked northeasterly trend, subparallel to the southeastern edge of the SRP and oblique to northwest-trending Quaternary normal faults in southeastern Idaho. The seismic belt continues northeasterly into western Wyoming to the vicinity of Jackson, immediately north of which there is a notable gap in seismicity coincident with the 70-km-long Teton fault. Intense seismicity occurs beneath the volcanically and hydrothermally active Yellowstone region and to its west in the Hebgen Lake region. A divergent belt of earthquake activity extends more than 400 km from Yellowstone Park in a west-southwest direction into central Idaho. This zone was originally described by Smith and Sbar (1974) as independent of the ISB and was termed the Idaho seismic zone (see also Smith, 1978). Stickney and Bartholomew (1987) similarly characterize it as an independent seismic zone, calling it the Centennial Tectonic Belt. This zone, however, forms part of an arcuate, parabolic pattern of seismicity flanking the SRP, with a vertex at Yellowstone Park that suggests causal influence by the Yellowstone-SRP (Y-SRP) volcanic system and related hot spot (Smith and others, 1985; Anders and others, 1989; Blackwell, 1989) and hence an integral relation with the main ISB.

The northern ISB (45¼° to 49°N) lies within the Northern Rocky Mountains province and extends more than 400 km in a northwest direction from Yellowstone Park to northwestern Montana, following a structural belt of Cenozoic basins bounded by Quaternary faulting of diverse trend. Earthquakes and west-northwest–striking faults between about 46½° and 48°N are interpreted by Stickney and Bartholomew (1987) to reflect, in part, an intraplate boundary called the Lewis and Clark Zone, which trends about N70°W through Missoula and Helena (Fig. 3).

SEISMOTECTONIC FRAMEWORK

Crustal structure and structural style

The geophysical framework of the Intermountain region has recently been summarized by Smith and others (1989). The crustal and upper-mantle velocity structure in the region is typified by the cross sections shown in Figures 4 and 5. The southern ISB coincides with a transition from thinner, extended crust and lithosphere on the west to thicker, more stable crust and

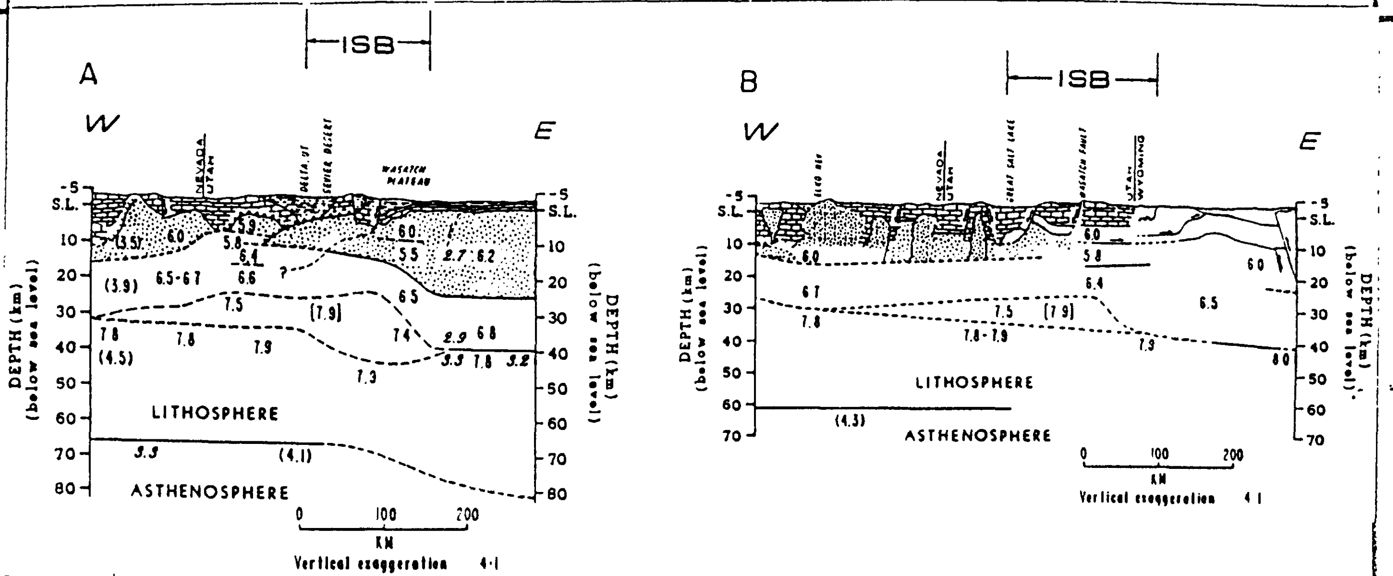

Figure 4. Cross sections of representative crustal and lithospheric structure across the southern Intermountain seismic belt (ISB), as summarized by Smith and others (1989). Profiles are roughly at latitude 38°N (A) and 41°-42°N (B); geographic reference features are shown in Figure 3. Superposed numbers indicate P-wave velocities (5.9-8.0 km/sec), S-wave velocities (3.5-4.5 km/sec, in parentheses), and densities (2.7-3.3 gm/cm^3, in italics); numbers in brackets are P_n-wave velocities corresponding to an alternative interpretation by Pakiser (1989) in which a mantle upwarp beneath the Basin and Range--Colorado Plateau transition has a uniform upper mantle velocity of 7.8 to 7.9 km/sec.

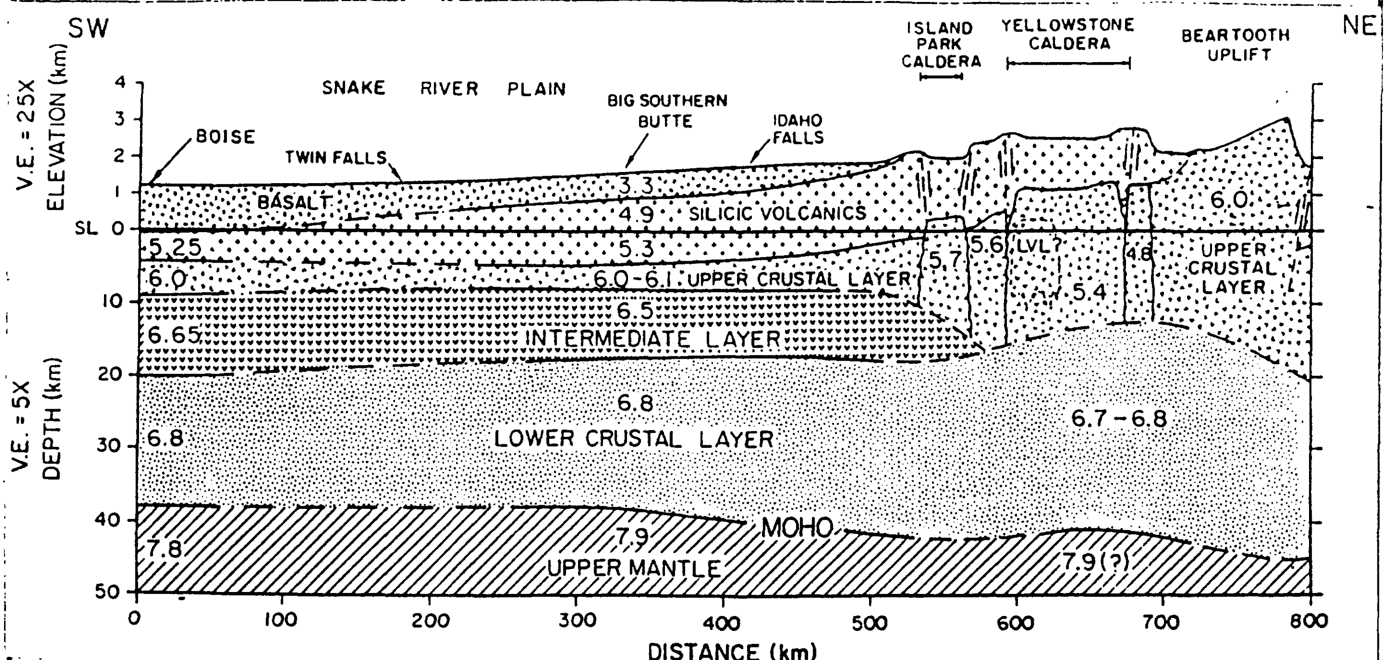

Figure 5. Cross section of crustal structure along the Snake River Plain in Idaho (representative of part of the central ISB) into the Beartooth uplift of Montana (representative of the northern ISB). Superposed numbers indicate P-wave velocities (km/sec). Data from 1978 and 1980 Yellowstone–Snake River Plain seismic-refraction experiment (Braile and others, 1982; Smith and others, 1982).

lithosphere on the east. The two profiles shown in Figure 4 extend from the central Basin and Range province in Nevada, where the Moho is about 30 km deep (below sea level) and the upper-mantle P_n velocity is 7.8 km/sec, to the Colorado Plateau-Middle Rocky Mountain provinces in Utah and Wyoming, where the Moho depth slightly exceeds 40 km and the P_n velocity reaches 8.0 km/sec. There is uncertainty about the depth of the true Moho beneath the transition region. If the top of a 7.8 to 7.9-km/sec layer marks the Moho beneath the Wasatch Front region of Utah (Fig. 4a), as identified by Loeb and Pechmann (1986), then the crust-mantle boundary is as deep as 45 km beneath the transition. Alternative interpretations imply the depth to the Moho may be as shallow as 25 km if observed velocities of 7.4 to 7.5 km/sec (Fig. 4) are the result of down-dip ray paths (see Smith and others, 1989, Fig. 4; Pakiser, 1989).

The P-wave velocity structure of the upper crust beneath the Y-SRP volcanic province, revealed by refraction/wide-angle reflection profiles, is more laterally heterogeneous than in the surrounding thermally undisturbed areas of the ISB. At Yellowstone a caldera-wide low-velocity body extends to depths of about 15 km and is thought to reflect a remnant magma reservoir with materials ranging from melts to hot, cooling granitic rocks (Smith and others, 1982). Along the SRP, the systematic decrease in elevation southwestward away from the caldera also reflects a systematic change in crustal structure. The near-surface basaltic layer thins northeastward from 2 km in southwestern Idaho, the suggested beginning of the trace of the Yellowstone hot spot, to zero thickness at Yellowstone; correspondingly, the deeper silicic layer thickens northeastward from zero thickness in southwestern Idaho to 2 km at the Yellowstone caldera, following the form of the surface topography (Fig. 5). A high-velocity, 6.5-km/sec layer cores the mid-crust of the eastern SRP and is interpreted as a solidified, mafic remnant of the crustal magma sources of the Yellowstone hot spot (Sparlin and others, 1982). This unusual high-velocity and high-density body may affect the overall strength and hence the seismic capability of the SRP. Note that neither the Moho nor the seismic velocity structure of the lower crust of the eastern SRP seem to have been altered by the youthful magmatism. The lower-crustal velocity structure beneath the SRP is the same as beneath adjacent thermally undisturbed regions.

We noted earlier that the ISB roughly follows the eastern margin of a broad domain of late Cenozoic extension in western North America (e.g., Eaton, 1982, Fig. 1). This margin also marks a thermal transition. The background heat flow is about 85 mW m^{-2} in the Basin and Range province and more than 100 mW m^{-2} in the SRP, contrasting with background values of less than 65 mW m^{-2} in the Colorado Plateau and in the area east of the Northern Rocky Mountains (Bodell and Chapman, 1982; Morgan and Gosnold, 1989). Locally, heat flow exceeds 1500 mW m^{-2} in the Yellowstone caldera (Blackwell, 1989). The ISB thus follows a structural and thermal transition to more stable continental interior with lower heat flow. The transition may be a locus of active lithospheric thinning (see Fig. 4A), occurring in general along the eastern margin of a regional-scale thermotectonic anomaly in the upper mantle. Effects of an active mantle hot spot associated with the Y-SRP system (discussed in a later section) are a special case.

Seismic-reflection profiling across Tertiary-Quaternary basins and active fault zones in the Basin and Range province has provided important information on upper-crustal structural style. The superposition of basin-range faulting upon pre-Neogene thrust belt structure, especially along the eastern margin of the Basin and Range province, is well known to be a fundamental and complicating factor (e.g., Smith and Bruhn, 1984). Anderson and others (1983), Allmendinger and others (1983), Smith and Bruhn (1984), and Smith and others (1989) have interpreted seismic-reflection data from throughout the northern Basin and Range province, identifying three characteristic styles of extensional basin development in the region: (1) relatively simple basins bounded by one or more planar normal faults dipping 45° to 60°; (2) asymmetric tilted basins displaced chiefly by a listric or planar low-angle normal fault; and (3) complex basins, typically with subbasins, associated with both planar and listric normal faults that sole into low-angle detachments.

Low-angle detachment faulting may have contributed substantially to Cenozoic crustal extension in at least part of the Intermountain region. The best example is the case of the Sevier Desert detachment (see sawteeth pattern beneath the Sevier Desert in Fig. 4A). High-quality seismic-reflection data (Allmendinger and others, 1983; Smith and Bruhn, 1984; Planke and Smith, 1991) show that the Sevier Desert detachment extends laterally with an average 10° to 15° westward dip at least 70 km from near the surface in central Utah to a depth of about 15 km beneath western Utah. The detachment has an estimated total surface area of 5,600 to 9,100 km^2 (Planke and Smith, 1991) and may have accommodated 30 to 60 km of Cenozoic extensional displacement (Allmendinger and others, 1983). In the Sevier Desert region, where the detachment lies only 3 to 5 km below the surface, prominent normal faults in the hanging wall do not cut the detachment but either abut or merge with it (Crone and Harding, 1984; Planke and Smith, 1991). The configuration of a high-angle normal fault, with Holocene(?) surface displacement, "directly connected" to the Sevier Desert detachment at relatively shallow depth, led Crone and Harding (1984) to raise concern about the detachment's seismogenic potential. Reviews of moderate to large normal-faulting earthquakes in diverse extensional regimes (Jackson and White, 1989; Doser and Smith, 1989) suggest that the low angle of dip of the Sevier Desert detachment make it an unlikely source of seismic slip, but the possibility of aseismic motion on the detachment cannot be ruled out.

There remain many uncertainties about the subsurface geometry of seismically active normal faults in the Intermountain region and whether seismic slip can occur on low-angle or listric normal faults known to be present. Moderately- to steeply-dipping planar geometries for the seismically active faults are typically inferred from aftershock locations and from the focal mechanisms of large earthquakes (e.g., Smith and others, 1985), a point we pursue later in this paper. Nevertheless, seismic-reflection evidence suggests that some segments of major normal faults with late Quaternary surface ruptures indeed have a listric subsurface geometry (Smith and Bruhn, 1984; Smith and others, 1989).

Throughout this paper we emphasize that the late Cenozoic structural style of the Intermountain region is dominated by normal faulting. There is growing awareness of the importance of strike-slip faulting as part of Neogene extensional deformation in the Basin and Range Province (Anderson, 1989), particularly in parts of the southern and western Great Basin (Rogers and others, this volume). In the Intermountain region, the best geologic evidence for Neogene strike-slip deformation is in the Sevier Valley area of south-central Utah (near Richfield, Fig. 3; Anderson and Barnhard, 1987). Focal mechanisms of background earthquakes in that same area also imply strike-slip faulting (Arabasz and Julander, 1986). Strike-slip focal mechanisms have also been observed for historical moderate-sized earthquakes in the northern ISB of Montana (Doser, 1989a), in the Hansel Valley area of northwestern Utah (near HV, Fig. 3; Doser, 1989b), and in southeastern Nevada (Smith and Sbar, 1974). Our present understanding of extension in the Intermountain region may underestimate the importance of strike-slip deformation.

Contemporary deformation

Regional stress field. Stress observations give important information about the pattern and mechanics of intraplate extension in the Intermountain region. Figure 6 shows the orientations of minimum horizontal compressive stress, S_{hmin}, variously deduced from the T-axes of focal mechanisms of moderate to large earthquakes, mapped fault-displacement vectors associated with Quaternary and Holocene slip events, orientations of volcanic dikes, borehole deformation, and in-situ hydrofracture stress measurements. The data are from a recent compilation by Zoback and Zoback (1989) for the continental United States.

Focal mechanisms and geologic indicators are the main sources of available stress information for the ISB. Given the ambiguity of selecting the correct fault plane in focal mechanisms, the principal stress axis directions were determined by assuming the standard Coulomb failure criterion with a coefficient of internal fraction of zero. This constraint places the maximum and minimum principal stress directions (corresponding to the P-axes, for maximum compression, and T-axes, for minimum compression) at 45° to the nodal planes. Because the dominant mode of contemporary deformation in the Intermountain region is extension, the T-axes give a general indication of relative motion within areas of coherent intraplate deformation and are consistent indicators of the directions of the minimum compressive stress for this region.

The overall stress field of the ISB (Fig. 6) is generally characterized by NE-trending S_{hmin} orientations in the northern ISB and the western part of the central ISB and ENE-to-ESE-trending S_{hmin} orientations in the southern ISB and the eastern part of the central ISB. The coherence of these orientations within the ISB and surrounding areas of the Rocky Mountains has led Zoback and Zoback (1989) to define a "Cordilleran extensional" province, larger than the traditional domain of the Basin and Range province and the Rio Grande rift. They distinguish the interior of the Colorado Plateau, however, as a distinct stress province with a WNW-trending orientation of maximum horizontal compressive stress S_{Hmax}. This is reflected in Figure 6 by the difference in

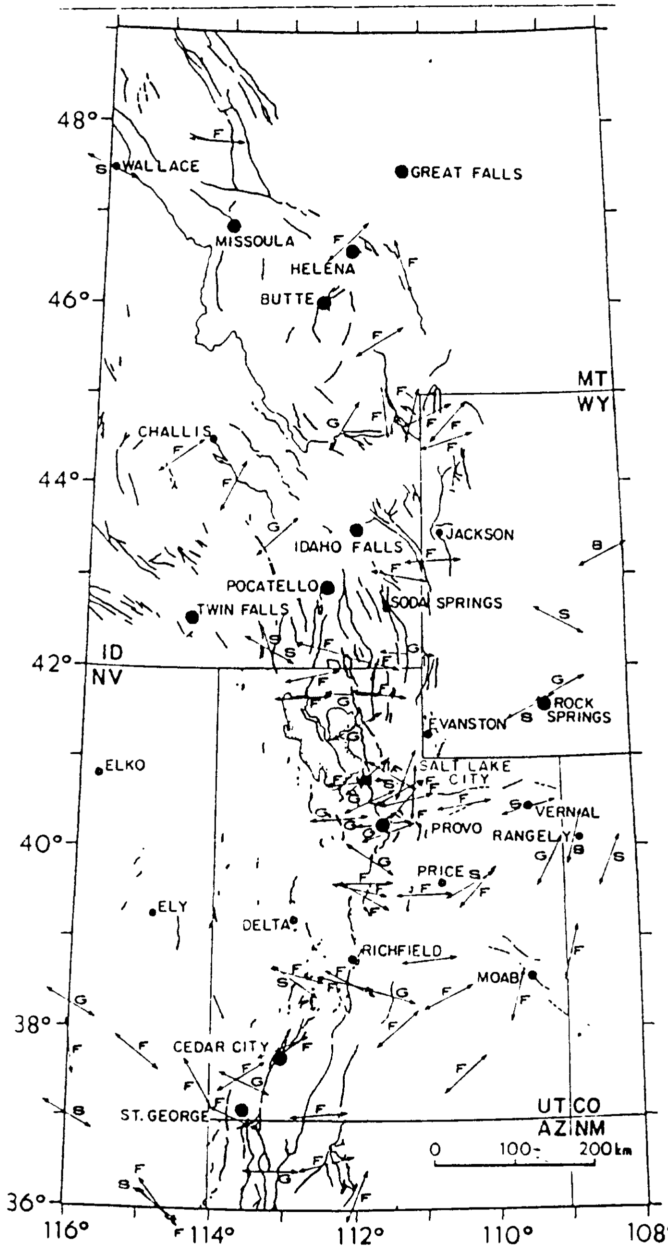

Figure 6. Map showing orientations of minimum horizontal compressive stress in the Intermountain region from a data compilation of Zoback and Zoback (1989), together with selected faults, as in Figure 3. Types of stress indicators are indicated as follows: B = borehole "breakout", F = focal mechanism, G = geologic, M = mixed, and S = in situ stress.

S_{hmin} orientations in eastern Utah and western Colorado from those along the main ISB to the west.

On the basis of focal-mechanism studies of small to large earthquakes in central Idaho and southern Montana, Smith and others (1977) and Stickney and Bartholomew (1987) distinguished a stress regime with N-trending S_{hmin} in the Hebgen Lake region, west of Yellowstone National Park, from the broader region of basin-range tectonism in the vicinity of the Montana-Idaho border where S_{hmin} is northeast-trending. Among other deviations from this pattern of northeast-trending S_{hmin} is a data point east of Helena for the 1925 Clarkston Valley earthquake of M_W = 6.6 (discussed below). This earthquake may have been more closely related to a stress field with northeast-southwest maximum horizontal compressive stress, characteristic of the more stable interior and thus transitional between the Basin and Range and the Great Plains (Doser, 1989a). We will comment on the stress implications of other sizable earthquakes in later sections.

An indicator of the relative magnitudes of the maximum (S_1), intermediate (S_2), and minimum (S_3) principal stresses has been defined by Bott (1959) as $\Phi = S_2 - S_3/S_1 - S_3$. Zoback (1989) and others have used this parameter to interpret regional stress variations within the Basin and Range province. Assuming that fault slip occurs in the direction of the maximum shear stress on a fault plane, then the orientation and sense of slip is governed by Φ and by the orientation of the principal stresses. For a normal faulting stress regime such as the ISB (S_1 vertical), if $\Phi = 0$, then the two horizontal stresses are equal and the predicted deformation is pure dip-slip for any fault orientation. For the maximum value of $\Phi = 1$, the vertical and maximum horizontal stresses are equal, and the predicted deformation is oblique-normal slip, transitional to strike-slip, depending on fault orientation.

Zoback (1989) and Bjarnason and Pechmann (1989) have recently summarized information on Φ-values for the southern ISB, variously from fault-slip data, focal mechanisms, in-situ stress measurements, and well-bore breakouts. In northern Utah, where the observed mode of faulting from fault-slip information and focal mechanisms is predominantly normal dip-slip and where S_{hmin} has an average east-west trend, Φ is highly variable. Averaged values of Φ (Zoback, 1989; Bjarnason and Pechmann, 1989) range from low values (0.0 to 0.3) for in-situ stress measurements and well-bore breakouts through intermediate values (0.3 to 0.7) for late Quaternary fault-slip measurements to slightly higher values (0.5 to 0.9) for focal mechanisms. In south-central Utah along the Basin and Range–Colorado Plateau transition, there is an observed mixture of dip-slip and strike-slip deformation, both in focal mechanisms (Arabasz and Julander, 1986) and in fault slip (Anderson and Barnhard, 1987). The observations can be explained by a high Φ-value and local changes in the relative magnitudes of (near-equal) vertical and maximum horizontal principal stresses under a relatively constant east-trending S_{hmin} (see Zoback, 1989). Bjarnason and Pechmann (1989) and Zoback (1989) have independently calculated an average Φ-value of 0.8 for grouped focal mechanisms in this area. There is little information on Φ-values for the central and northern ISB, except for a study of the main shock and aftershocks of the 1983 Borah Peak, Idaho, earthquake by Smith and others (in preparation), who find a Φ-value of 0.65, consistent with the observed oblique-normal faulting (described below).

Strain rates. Earthquake focal mechanisms and seismic moments have been used by Eddington and others (1987) to calculate regionalized strain rates and corresponding deformation rates produced by historical earthquakes in the western United States, following the method of Kostrov (1974). Available results for the central and southern ISB are shown in Figure 7. The figure shows 14 selected areas of inferred homogeneous stress in which the moment tensors of historical earthquakes have been summed and diagonalized to get the direction and magnitude of horizontal principal strain. Together with the S_{hmin} orientations in Figure 6,

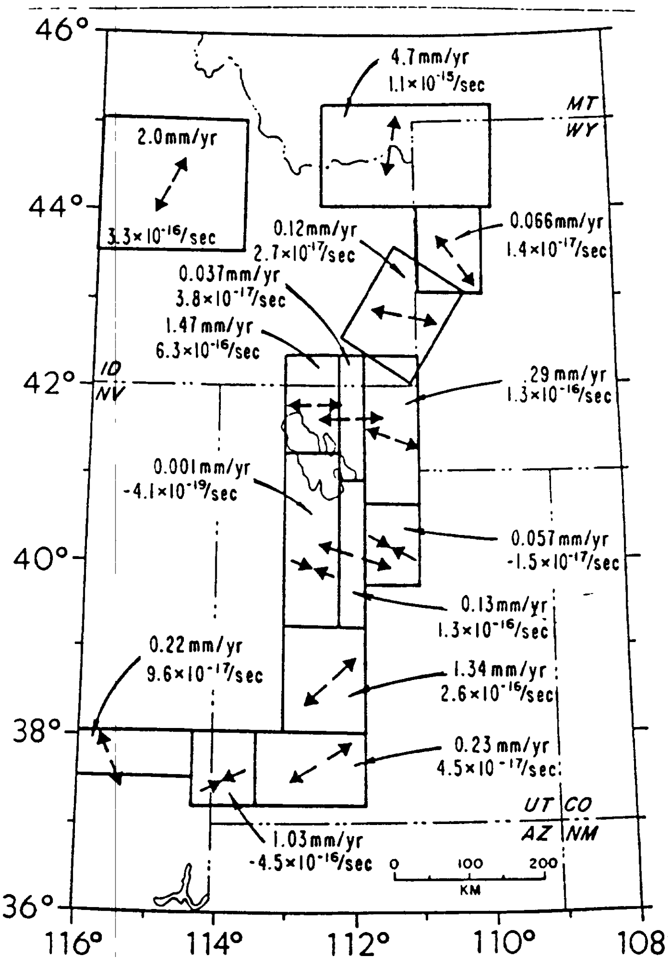

Figure 7. Strain and deformation rates in part of the Intermountain region based upon summations of seismic moments from historical earthquakes, after Eddington and others (1987). Earthquakes within areas of assumed homogeneous strain, shown by boxes, were used to determine the summed seismic moment. Information shown for each box includes: orientation of the horizontal principal strain axis (arrows); the deformation rate, in mm/yr (upper number); and the horizontal strain rate, in sec^{-1} (lower number).

the strain-rate and deformation data in Figure 7 provide an overall perspective of contemporary deformation in the central and southern ISB.

In the central ISB (Figure 7, top), historical seismicity in central Idaho implies NNE–SSW extensional strain and yields a strain rate of the order of 10^{-16}/sec. In the Hebgen Lake–Yellowstone Park region to the east, the most seismically active region of the entire U.S. Cordillera, the extensional strain rate of 1.1×10^{-15}/sec (4.7 mm/yr deformation rate) is more than three times greater, and the horizontal principal strain axis trends more northerly. For parts of the central ISB south of Yellowstone Park, much smaller extensional strain rates of the order of 10^{-17}/sec and deformation rates of 0.07 to 1.2 mm/yr were calculated. The northwest-southeast direction of horizontal principal strain in the Teton region is considered uncertain because of sparse data; in southeastern Idaho, a better-resolved extensional strain direction is nearly east-west.

In the southern ISB, general east-west extensional strain of the order of 10^{-16}/sec or smaller characterizes Utah's Wasatch Front area. Deformation rates range from 0.001 mm/yr in the western part of the Wasatch Front area to 1.5 mm/yr in the northwestern part. Regarding the latter subregion, we note that the depicted strain information would not be significantly affected by a recently revised focal mechanism (Doser, 1989b) implying predominantly strike slip rather than dip slip for the M = 6.6 Hansel Valley earthquake of 1934. Because the direction of the T-axis for the revised solution (Doser, 1989b) is nearly identical to that for the original dip-slip solution (Dewey and others, 1973) used by Eddington and others (1987), the averaged moment rate and direction of deformation given in Figure 7 would not be significantly changed. In southern Utah, extensional strain rates of 10^{-16}/sec to 10^{-17}/sec are comparable to those for most of the Wasatch Front area, but the strain direction is northeast-southwest. A possible change to northeast-southwest compression is suggested for one subregion in southwesternmost Utah. The significance of this local apparent change in mode of deformation is uncertain. The northwest-southeast extensional strain direction shown for a subregion in southeastern Nevada is consistent with that for other parts of southern and central Nevada analyzed by Eddington and others (1987).

Deformation rates intrinsically depend on the dimensions of the subregion being considered, so comparisons must be made with care. Nevertheless, data summarized by Eddington and others (1987) for the western United States make it evident that deformation rates for most of the intraplate ISB are one to two orders of magnitude lower than those along the western North American plate boundary. The ISB deformation rates can be evaluated in another way. Eddington and others (1987) summed earthquake deformation rates along east-west profiles across the northern and southern Great Basin, including the ISB, to obtain integrated extension rates of 8 to 10 and 3 to 4 mm/yr, respectively, with approximately 1 mm/yr taken up across the ISB (Doser and Smith, 1982). Estimates of Late Cenozoic total deformation rates of 1 to 20 mm/yr for the Great Basin determined from other geologic and geophysical data (Table 1) are of the same magnitude as the contemporary earthquake-induced rates.

TABLE 1. RATES OF EXTENSION ACROSS THE GREAT BASIN *

Time	Deformation Rate (mm/yr)	Method
Late Cenozoic	3 - 20	Geological strain
Late Cenozoic	3 - 12	Heat flow
Holocene paleoseismicity	1 - 12	Fault-slip data
Historic seismicity	3.5 - 10	Historical earthquakes
Inferred Quaternary	<9	Intraplate models

* Adapted from Eddington and others (1987, Table 5); sources identified therein.

Quaternary faulting

To compare the regional seismicity of the ISB with active faulting, a generalized map was made of late Tertiary to Holocene normal faulting in the Intermountain region (Fig. 3) from the following sources. The locations of active fault traces compiled by Witkind (1975a, b, c) were digitized for Idaho (except for central Idaho), western Wyoming, and western Montana. Surface traces of late Quaternary faults of central Idaho and southwestern Montana, including the surface rupture of the 1983 Borah Peak earthquake (M_S = 7.3), were taken from compilations by Haller (1988) and Scott and others (1985). (A more detailed map of active faulting in the Y-SRP area is presented in Fig. 15, below.) Fault data for the Utah region are from a compilation of Arabasz and others (1987) for the Wasatch Front area (outlined in Fig. 3), supplemented elsewhere by data from Anderson and Miller's (1979) Quaternary fault map of Utah. Quaternary faults for northern Arizona are from the Arizona Quaternary fault map of Scarborough and others (1986). We did not attempt to make a complete compilation of fault data for eastern Nevada as this area was considered marginal to our discussion. Figure 3 thus serves as a fair representation of known or suspected active faults in late Cenozoic time in the Intermountain region, but we caution that the fault compilation is non-uniformly complete.

Age of normal faulting and correlation with topography. Dating the inception and evolution of normal faulting on individual faults in the Intermountain region is generally handicapped by a lack of suitable exposures and datable materials. There are some exceptions. Geochemical studies of altered fault rock in the exhumed footwall of the Wasatch fault (Parry and Bruhn, 1986, 1987) indicate an origin at 11 km depth 17.6 ± 0.7 m.y. ago. The Teton fault in Wyoming has a total vertical displacement of as much as 6 to 9 km that began 7 to 9 m.y. ago (Love and Reed, 1971). On a regional basis, there is evidence for two stages of Cenozoic extensional tectonism and normal faulting in many parts of the Basin and Range province. As reviewed by Levy and Christie-Blick (1989) and Anderson (1989), a first early stage of Cenozoic extension began about 37 Ma and was apparently restricted to a relatively narrow region of high strain in eastern Nevada, western Utah, and southern Idaho and was accompanied by calc-alkaline volcanism, detachment faulting, and core-complex formation. The second stage was the classic episode of Basin and Range extension and epeirogeny responsible for the present topography and the steeply dipping normal faults of the ISB. This modern stage generally began 15 to 10 Ma in the northern Basin and Range province, but earlier in the southern part of the province.

The distinctive north-northeast- to northwest-trending basin-range topography of the ISB involves sediment-filled basins and tilted range blocks bounded by large normal faults with significant Quaternary displacement. The observed coseismic deformation associated with normal faulting during large earthquakes and theoretical modeling suggest that both footwall uplift of the mountain blocks and hanging-wall subsidence of the adjacent asymmetric basins have been fundamental in developing basin-range topography (e.g., King and others, 1988; see also Fig. 24, below). For individual ranges, relative crest height along the range seems to correlate with the size and frequency of young fault displacements along the base of the range. For example, along the Teton fault in Wyoming, the maximum heights of Quaternary scarps ranging up to 50 m high correspond with the highest parts of the Teton Range (Smith and others, 1990a, b). The 1983 Borah Peak earthquake (M_S = 7.3) ruptured one of the most active central segments of the Lost River fault zone, which was coincidentally adjacent to the highest part of the Lost River Range (Scott and others, 1985), including Borah Peak, the highest point in Idaho. Crest heights along the Wasatch Range are greater along the active central segments of the Wasatch fault than along distal segments with lower late Pleistocene–Holocene slip rates (Schwartz and Coppersmith, 1984). Relative topographic relief serves as an indicative but insufficient guide to the location of segments of range-front normal faults likely to produce future surface-faulting earthquakes.

Threshold of surface faulting and maximum magnitude

Summaries of information and discussion about the minimum magnitude needed to produce coseismic surface faulting in the Intermountain region are given by Doser (1985a) and Arabasz and others (1987). The threshold magnitude appears to be in the range of $6.0 \leqslant M_L \leqslant 6.5$, based on the historical record of earthquakes in the ISB and in the Basin and Range province. Arabasz and others (1987) adopt $M_L = 6.3 \pm 0.2$ as an estimate of the threshold in the Utah region and argue that earthquakes up to this size can occur anywhere in the southern ISB, even where there is no geologic evidence for Quaternary surface faulting.

The M_S = 7.5 Hebgen Lake earthquake of 1959 (described in detail herein) is considered by some to represent the maximum earthquake size for the ISB (e.g., Doser, 1985a). This earthquake, however, was smaller than at least two other large earthquakes in the Basin and Range province. The 1872 Owens Valley, California, earthquake had an estimated moment magnitude of 7¾ to 8 and was associated with a predominantly strike-slip surface rupture up to 110 km long, with a maximum lateral offset of 7 m and a maximum vertical offset of 4.4 m; the 1915 Pleasant Valley, Nevada, earthquake of surface-wave magnitude 7.6 had a 60-km-long rupture and a maximum vertical displacement of 5.8 m (dePolo and others, 1989). The Hebgen Lake earthquake had a shorter rupture length, but had comparable displacement to these two earthquakes. Because there are adjoining fault segments in the ISB with potential rupture lengths exceeding that of the Hebgen Lake earthquake, and comparable to those of the Pleasant Valley and Owens Valley earthquakes, it is reasonable to consider maximum-magnitude earthquakes slightly greater than M_S = 7.5 for particular faults in the ISB (e.g., Arabasz and others, 1987, adopt a maximum magnitude of Ms = 7.5 to 7.7 for the Wasatch Front area).

Estimates of the maximum magnitude for earthquakes on any particular fault in the ISB have uncertainties relating not only to the prediction of future rupture characteristics but also to the conversion of those rupture characteristics to estimated magnitudes. This problem is particularly important for seismic hazard analysis. For example, in an evaluation of probabilistic ground shaking for the Wasatch Front, Youngs and others (1987) used fault length-magnitude relations of Bonilla and others (1984) to estimate maximum magnitudes of M_S = 7.2 to 7.5 for the longest segments of the Wasatch fault in northern Utah. In a seismotectonic study of the Jackson Lake dam in Wyoming, Gilbert and others (1983) selected a maximum credible earthquake of magnitude (unspecified scale) 7.5 based on comparisons of observed scarp lengths and fault length with rupture length-magnitude relations such as those of Slemmons (1977). Thenhaus and Wentworth (1982) suggested a maximum magnitude of 7½ for eastern Idaho and central and western Utah, and adopted a value of 7¾ specifically for the Wasatch fault. For a site at the Idaho National Engineering Laboratory near Idaho Falls, Idaho, Woodward-Clyde Consultants (1979) relied upon a global compilation of data for surface-faulting earthquakes to assign a peak ground acceleration, using the 1959, M_S = 7.5 Hebgen Lake earthquake as a maximum-magnitude event for that area.

THE EARTHQUAKE RECORD IN THE INTERMOUNTAIN REGION

The traditional earthquake record for the Intermountain region—consisting of historical seismicity (based on non-instrumental reports of felt earthquakes) and instrumental seismicity—is much less substantial than for other parts of the United States. This is because of the region's relatively late settlement, historically sparse population, and late seismographic coverage. In contrast, comparatively abundant information has been gathered in this part of the United States on the timing and character of prehistoric earthquakes from paleoseismology. In this section, we restrict attention to the traditional earthquake record.

Pre-instrumental information

Reports of historical seismicity in the Intermountain region date only from the mid 1800s when systematic modern settlement began. Diverse Indian cultures were well established in the region after A.D. 100 to 1300 but left no known record of specific felt earthquakes. During the 1820s and 1830s, a few hundred fur trappers made up most of the non-Indian population of the region. Mormon pioneers entered the Salt Lake Valley in 1847 and promptly started a regional colonizing program. Within a few decades, Mormon settlements extended throughout large portions of the Intermountain West (Wahlquist, 1981), with the exception of Montana, where the settlements began chiefly as mining camps after the 1850s. The first documented earthquakes in the Intermountain region date from 1850 in Utah (Arabasz and McKee, 1979), 1869 in Montana (Qamar and Stickney, 1983), 1871 in Wyoming (Hayden, 1872), and 1879 in Idaho (Townley and Allen, 1939). The highly non-uniform distribution of population before 1900 throughout the Intermountain region implies great variability in the threshold of detection and in location errors for pre-instrumental earthquakes.

The pre-instrumental earthquake record for the Intermountain region comes from multiple sources. Coffman and others (1982) cite many early reports, records, and compilations for earthquakes in the "Western Mountain Region" before 1928. For 1928 and later, annual reports published by the U.S. Department of Commerce under the title *United States Earthquakes* are key sources. Williams and Tapper (1953) made an important historical study of Utah earthquakes from 1850, the time of publication of the first newspaper in Utah, through 1949. Cook and Smith (1967) extended this record to 1965, including computer determinations of the first instrumental seismicity for the Utah region from systematic regional recording. Modern compilations of historical seismicity in the study area include those of Arabasz and McKee (1979) and Stover and others (1986), for the Utah region, and Qamar and Stickney (1983) for Montana.

Instrumental recording and seismic networks

Seismographic recording in the Intermountain region began with the installation of two modified Bosch-Omori pendulum seismographs on the University of Utah campus in Salt Lake City in 1907 (Arabasz, 1979). Figure 8 shows stages of subsequent instrumental coverage of the Intermountain region in 1948, 1968, and 1988. By 1948, electromagnetic seismographs were operating in at least eight locations in the region (dated circles, Fig. 8A). Systematic reporting of seismological data to the U.S. Coast and Geodetic Survey (USCGS) from Salt Lake City began in 1938, one year before the Bosch-Omoris were replaced by more modern instruments (Arabasz, 1979). For each of the other dated stations shown in Figure 8A, reporting to the USCGS began immediately after installation (see *United States Earthquakes*).

In 1968, there were at least 25 seismographic stations in the study area (triangles, Fig. 8A), virtually all with on-site recording. The changes from 1948 to 1968 almost exclusively reflect additions in the 1960s (see Poppe, 1980). These relate to stations in the western part of the area shown in Figure 8A installed with the motivation, in part, to record underground nuclear explosions from the Nevada Test Site, the development in Utah of a skeletal statewide network (Arabasz and others, 1979), local monitoring of mining-related seismicity in east-central Utah and in northern Idaho, dam-site monitoring at Flaming Gorge and Glen Canyon (Lake Powell) (Fig. 3), and added USGS coverage of the Hebgen Lake–Yellowstone area. The 1968 "snapshot" misses the presence of scattered Department of Defense, LRSM mobile seismic observatories operated on a temporary basis in the mid 1960s at a few dozen sites throughout the Intermountain region (see Poppe, 1980). The time frame does include, however, operation of a Department of Defense, VELA–Uniform array at the Uinta Basin Observatory (UBO) in northeastern Utah. Figure 8A also includes WWSSN stations installed in 1962 at Dugway (DUG), Utah, and in 1963 west of Bozeman (BOZ), Montana. The latter station operated until 1968 and was moved to Missoula, Montana, in 1973.

By the mid to late 1970s, short-period seismic telemetry networks had become well established in the Intermountain region. Seismographic coverage of the region shown for 1988 (Fig. 8B), with a total of nearly 150 stations, is chiefly a composite of three regional and five local networks (see caption for Fig. 8B). Representative reporting and details for some of the network monitoring are given by Stickney (1988) for Montana, Peyton and Smith (1990) for Yellowstone, King and others (1987) for eastern Idaho, Wood (1988) for western Wyoming–eastern Idaho, Nava and others (1990) for the Utah region, and Rogers and others (1987) for southern Nevada. (See also Wong and Humphrey, 1989, regarding local network monitoring within the Colorado Plateau of southeastern Utah between 1979 and 1987.)

Earthquake catalog

In this chapter we use the catalog through 1985 of Engdahl and Rinehart (1988; this volume), compiled for the 1988 Seismicity Map of North America and hereafter referred to as the DNAG catalog, in order to present an overview of the whole Intermountain region. Original pre-instrumental data are chiefly from sources already described in a preceding section. Sources of instrumental data vary with time, depending on the evolution of seismographic coverage in the region. Instrumental locations for sizable earthquakes in the Intermountain region were made in the

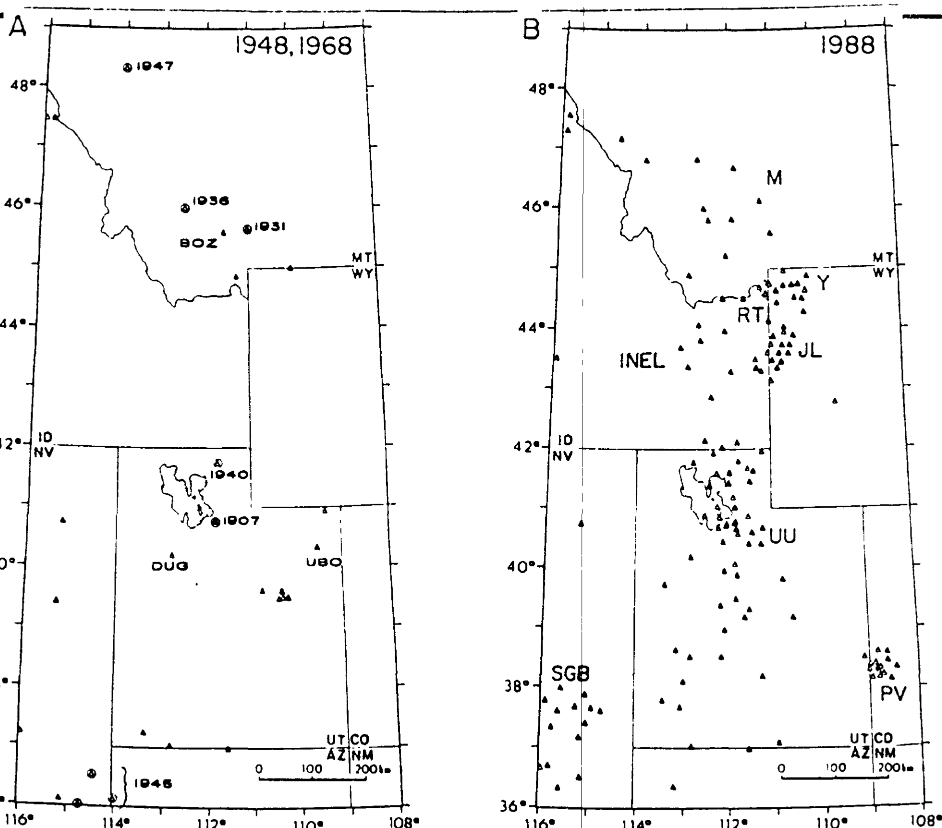

Figure 8. Maps showing representative distribution of seismographic stations in the Intermountain region at three selected times. A, Stations operating in 1948 (circles dated with year of installation) and in 1968 (triangles); circumscribed triangles, both 1948 and 1968. Key: 1907 = Salt Lake City; 1931 = Bozeman; 1936 = Butte; 1940 = Logan; 1945 = Boulder City, Overton, and Pierce Ferry (all surrounding Lake Mead). B, Stations operating in 1988, chiefly as parts of separate seismic networks. Key to seismic networks (from north to south): M = Montana (Montana Bureau of Mines and Geology, 12 sta., from ca. 1980); Y = Yellowstone (U.S. Geological Survey and University of Utah, 16 sta., from 1973); RT = Ricks-Teton (Ricks College, 5 sta., from 1972); JL = Jackson Lake (U.S. Bureau of Reclamation, 16 sta., from 1985); INEL = Idaho National Engineering Laboratory (6 stations, from ca. 1972; UU = University of Utah (57 sta. operated, 82 sta. recorded, from 1974); PV = Paradox Valley (U.S. Bureau of Reclamation, 15 sta., from 1983); SGB = Southern Great Basin (U.S. Geological Survey, 54 sta. [not all within figure], from 1978).

1920s to 1940s by the California Institute of Technology (Gutenberg and Richter, 1954) and routinely after the 1930s by the USCGS and later the U.S. Geological Survey (see *United States Earthquakes*). Figure 8A makes it evident that instrumental locations before the 1960s had to be based on recordings at widely spaced stations in the western U.S.

In mid-1962, the University of Utah began regional instrumental monitoring that later allowed compilation of an important catalog for the Utah region (36.75° to 42.50°N, 108.75° to 114.25°W), predating the installation of a modern (telemetered) regional network in 1974 (Arabasz and others, 1980). These data and subsequent, modern regional-network data make up primary sources of instrumental information in the DNAG catalog for the region (Engdahl and Rinehart, 1988). The latter include data from the University of Utah, the Montana Bureau of Mines and Geology, the Idaho National Engineering Laboratory, and the U.S. Geological Survey (Yellowstone National Park and southern Nevada). Other relevant data sources and compilations are described by Eddington and others (1987).

Time-varying thresholds of completeness since 1900 in the DNAG catalog are suggested by Engdahl and Rinehart (1988; this volume). For the study area, the overall record is best for the Utah region, where the threshold has been about magnitude 2.5 (3.0 in some distal areas) since 1962 and about magnitude 5¾ (Modified Mercalli Intensity VII) for perhaps the entire historic record (Rogers and others, 1976). For the region as a whole, our own subjective judgment indicates catalog completeness above magnitude 5¾ since 1900, above magnitude 5.0 since the 1920s to 1930s, and above magnitude 4.0 since the 1960s. Thresholds of completeness at and below magnitude 2.5 are associated with the modern regional network recording, but current completeness for the entire Intermountain region based on that recording (Fig. 8B) is at about magnitude 3.0.

The precision of instrumental earthquake locations in the region varies considerably in time and space (see original sources of data). Revised epicenters for instrumentally recorded earthquakes before the 1960s generally have uncertainties of tens of kilometers (e.g., Dewey and others, 1973; Qamar and Hawley, 1979). For earthquakes since the 1960s, epicentral precision reaches ±2 km or better within areas where seismographic spacing is of the order of a few tens of kilometers—as for some of the modern networks (Fig. 8B). Where the station spacing becomes greater, epicentral uncertainties are typically ±5 km, commonly increasing to ±10 km for events outside or in a distal part of the recording network. Reliable focal depths, requiring the presence of a recording station within roughly one focal depth of an earthquake's epicenter, are available for only a small fraction of the earthquakes in the DNAG catalog for the region.

Largest historical earthquakes

For the period from 1900 through 1985, the DNAG catalog contains 49 earthquakes in the Intermountain region with an indicated magnitude of 5.5 or greater, a selected threshold related to the potential for seriously damaging ground motions. These earthquakes are listed in Table 2 and plotted in Figure 9. The conventional magnitudes given in Table 2 are representative estimates, not necessarily the values listed in the DNAG catalog. Locations plotted in Figure 9 are directly from the DNAG catalog; refined locations, where available, are substituted in Table 2. In the following subsections we sequentially describe: (1) two known historical shocks of estimated magnitude 5.5 or greater from the pre-1900 period; (2) the four largest earthquakes in the Intermountain region's recorded history—all later than 1900, all with moment magnitudes greater than 6.5, and all but one with associated surface faulting (Table 2); and (3) other significant earthquakes after 1900 within each of the three main parts of the ISB. The descriptions are necessarily abbreviated, commonly citing one or more relevant summaries in place of original sources. For brevity, MMI signifies Modified Mercalli intensity.

Pre-1900 period. Two significant earthquakes occurred in the Intermountain region during the pre-1900 historical period. An earthquake on November 9, 1884, at 02:00 (local time) was felt strongly in Idaho, Utah, and Wyoming over at least 15,000 km^2 (Williams and Tapper, 1953). Descriptions of damage, MMI = VIII, and reports of at least six shocks felt at Paris, Idaho, in the Bear Lake Valley led Arabasz and McKee (1979) to assign an epicenter at 42.0°N, 111.3°W, arbitrarily on the Idaho-Utah border astride the active East Bear Lake fault, and to estimate a magnitude of 6.3, assuming a relation between MMI and magnitude from Gutenberg and Richter (Richter, 1958). A magnitude of at least 5½ seems likely. On November 4, 1897, at 02:29 (local time) a sizable earthquake occurred in southwestern Montana, with an assigned location of 45.0°N, 113°W, causing damage and resulting in MMI = VI at Dillon, Montana (Coffman and others, 1982). Estimating a felt area of about 500,000 km^2, Qamar and Stickney (1983) assigned a magnitude of 6.4, using an empirical relation between magnitude and felt area for Montana earthquakes.

1925 Clarkston Valley, Montana, earthquake. This 1925 earthquake (No. 11, Table 2), the second largest historical earthquake in Montana, occurred about 50 km northwest of Bozeman in the vicinity of Clarkston Valley, a late Cenozoic intermontane basin bounded on the east by the Clarkston Valley normal fault (Qamar and Hawley, 1979). Despite its significant size (M_{GR} = 6¾, M_W = 6.6, Table 2), the earthquake apparently produced no primary surface faulting, although ground cracks were observed at several localities (Pardee, 1926). The earthquake reached MMI = VIII, was felt over 800,000 km^2, caused major rockfalls, and resulted in considerable damage at nearby towns (Coffman and others, 1982; Qamar and Stickney, 1983). Seismologic data for the main shock (Doser, 1989a; Doser and Smith, 1989) indicate a mainshock focal depth of about 9 km, a subsurface rupture length of about 12 km, and oblique normal slip on a northwesterly-dipping plane with an orientation similar to that of the southern end of the Clarkston Valley fault. The main shock was preceded by at least one sizable foreshock and was followed by aftershocks as large as magnitude 4.8 (Doser, 1990).

1934 Hansel Valley, Utah, earthquake. The 1934 Hansel Valley (Kosmo) earthquake (No. 15, Table 2) remains the largest earthquake in the Utah region since 1850 and the only historical shock in the southern ISB known to have produced surface faulting. The earthquake occurred about 60 km west of the Wasatch fault in a sparsely populated, basin-range setting at the northern end of the Great Salt Lake (Fig. 3-2). The earthquake was felt over an area of 440,000 km^2 and reached MMI = VIII; it caused relatively minor damage (Coffman and others, 1982), although two deaths resulted, one direct and one indirect (Cook, 1972). Richter (1935) used the earthquake as an example in defining his magnitude scale, assigning a magnitude of 7.0 (probably overestimated because of an uncertain distance correction to Pasadena). The conventional magnitude of 6.6 assigned by Gutenberg and Richter (1954) is apparently a surface-wave magnitude and is identical to the earthquake's modern moment magnitude, M_W, of 6.6 calculated by Doser (1989b).

Shenon (1936) provided key documentation of geologic effects of the earthquake, including surface ruptures, rock slides, liquefaction, and other ground-water effects (see Arabasz, 1979, Doser, 1989b, and dePolo and others, 1989, for reference to other original sources). Shenon (1934) mapped four northerly-trending subparallel fractures displacing salt flats and unconsolidated late Quaternary sediments in the southwestern part of Hansel Valley over a zone about 6 km wide and 12 km long (see Doser, 1989b). Displacements were primarily vertical, up to a maximum of 50 cm, but a horizontal offset of 25 cm was also reported (dePolo and others, 1989). The relation of the 1934 surface rupturing to local geologic structure and neotectonics is of

TABLE 2. EARTHQUAKES IN THE INTERMOUNTAIN SEISMIC BELT OF MAGNITUDE 5.5 AND GREATER, 1900 THROUGH 1985

No.	Date (GMT)		Time (GMT) hr mn	Lat. (°N)	Long. (°W)	Magnitude M_{conv}		M_W	Region	References
1.	1900	Aug 01	07:45	40.0	112.1	$(5\frac{1}{2}\pm)$	I/A	—	Eureka, Utah	1, 2, 3L
2.	1901	Nov 14	04:39	38.8	112.1	$(6\frac{1}{2}\pm)$	I/A	—	Southern Utah (Richfield)	1, 2, 3L
3.	1902	Nov 17	19:50	37.4	113.5	$(6\pm)$	I/A	—	Pine Valley, Utah	1, 2, 3L
4.	1905	Nov 11	21:26	42.9	114.5	$(5\frac{1}{2}\pm)$	I/A	—	South central Idaho (Shoshone)	2L, 4
5.	1909	Oct 06	02:50	41.8	112.7	$(6\pm)$	I/A	—	NW Utah (Hansel Valley)	1, 2, 3L
6.	1910	May 22	14:28	40.8	111.9	$(5\frac{1}{2}\pm)$	I/A	—	Salt Lake City, Utah	1, 2, 3L
7.	1912	Aug 18	21:12	36.5	111.5	$(5\frac{1}{2}\pm)$	I/A	—	NE of Williams, Ariz.	2L
8.	1914	May 13	17:15	41.2	112.0	$(5\frac{1}{2}\pm)$	I/A	—	Ogden, Utah	1, 2, 3L
9.	1921	Sep 29	14:12	38.7	112.2	$(6\pm)$	I/A	—	Elsinore, Utah	1, 2, 3L
10.	1921	Oct 01	15:32	38.7	112.2	$(6\pm)$	I/A	—	Elsinore, Utah, 2nd main shock	1, 2, 3L
11.	1925	Jun 28	01:21	46.00	111.50	$6\frac{3}{4}$	M_{GR}	6.6	Clarkston Valley, Mont.	(8, 12)L, (5, 6)S
12.	1928	Feb 29	22:38	46.6	112.0	$(5\frac{1}{2}\pm)$?	I/A	—	Helena, Mont.	7LS, 8S
13.	1929	Feb 16	03:00	46.1	111.3	5.6	I/A	—	Lombard, Mont.	8LS
14.	1930	Jun 12	09:15	42.6	111.0	5.8	REN	—	Grover, Wyo.	2L, 10S
15.	**1934**	**Mar 12**	**15:05**	**41.77**	**112.67**	6.6	M_{GR}	6.6	Hansel Valley, Utah	6S, 9LS
16.	1934	Mar 12	18:20	41.57	112.75	6	M_{GR}	5.9	Hansel Valley, Utah, aftershock	6S, 9LS
17.	1934	Apr 07	02:16	41.5	111.5 ?	5.5	REN	—	Hansel Valley, Utah,aftershock?	1, 7L, 10S
18.	1934	Apr 14	21:26	41.73	112.60	5.6	REN	—	Hansel Valley, Utah, aftershock	9L, 10S
19.	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	$5\frac{1}{2}$	M_{GR}	—	Hansel Valley, Utah, aftershock	6S, 9L
20.	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	I/A	—	Helena, Mont., swarm event	[illegible]
21.	1935	Oct 19	04:48	46.80	112.[illegible]	$6\frac{1}{4}$	M_{GR}	6.2 ±	Helena, Mont., swarm event	5LS, 6S
22.	1935	Oct 31	18:37	46.62	111.97	6	M_{GR}	6.0 ±	Helena, Mont., swarm event	5LS, 6S
23.	1935	Nov 28	14:41	46.60	112.00	5.5	REN	—	Helena, Mont., swarm event	2L, 10S
24.	1936	May 13	14:06	46.60	112.00	5.7	?	—	Helena, Mont., swarm event	7LS, 8
25.	1936	May 22	02:19	46.60	112.00	5.7	?	—	Helena, Mont., swarm event	7LS, 8
26.	1944	Jul 12	19:30	44.41	115.06	6.1	PAS	—	Central Idaho (Seafoam)	2, 13LS
27.	1945	Feb 14	03:01	44.61	115.09	6.0	PAS	—	Central Idaho (Clayton ?)	2, 13LS
28.	1945	Sep 23	09:57	48.00	114.20	5.5	I/A	—	Flathead Lake, Mont.	2L, 8S
29.	1947	Nov 23	09:46	44.92	111.53	$6\frac{1}{4}$	M_{GR}	6.1	Virginia City, Mont.	6S, 5LS
30.	1952	Apr 01	00:37	48.00	113.80	5.5	I/A	—	Big Fork, Mont.	2L, 8S

TABLE 2. EARTHQUAKES IN THE INTERMOUNTAIN SEISMIC BELT OF MAGNITUDE 5.5 AND GREATER, 1900 THROUGH 1985 (continued)

No.	Date (GMT)		Time (GMT) hr mn	Lat. (°N)	Long. (°W)	Magnitude M_{conv}		M_w	Region	References
31.	1959	Jul 21	17:39	37.00	112.50	5.7	PAS	—	Arizona-Utah border	2L, 7S
32a.	1959	Aug 18	06:37	44.88	111.11	6.3	m_b	6.3	Hebgen Lake, Mont., double event	14LS, 15S
32b.	**1959**	**Aug 18**	**06:37**	**44.84**	**111.03**	7.5	M_S	7.3	Hebgen Lake, Mont., double event	14L, (15,16)S
33.	1959	Aug 18	07:56	45.00	110.70	$6\frac{1}{2}$	BRK	—	Hebgen Lake, Mont., aftershock	15L, 20S
34.	1959	Aug 18	08:41	45.08	111.80	6	BRK	—	Hebgen Lake, Mont., aftershock	15L, 20S
35.	1959	Aug 18	11:03	44.94	111.80	$5\frac{1}{2}+$	BRK	—	Hebgen Lake, Mont., aftershock	15L, 20S
36.	1959	Aug 18	15:26	44.85	110.70	6.3	M_S	6.3	Hebgen Lake, Mont., aftershock	15LS
37.	1959	Aug 19	04:04	44.76	111.62	6	BRK	6.0	Hebgen Lake, Mont., aftershock	5L, (5, 20)S
38.	1962	Aug 30	13:35	41.92	111.63	5.7	M_S	5.6	Cache Valley (Logan), Utah	17LS
39.	1964	Oct 21	07:38	44.86	111.60	5.8	m_b	5.6	Hebgen Lake, Mont., aftershock	5LS, 7S
40.	1966	Aug 16	18:02	37.46	114.20	6.0	PAS	5.3	Southeast Nevada	7S, 15LS
41.	1966	Aug 18	10:09	37.30	114.20	5.6	m_b	—	Southeast Nevada, aftershock	7LS
42.	1975	Mar 28	02:31	42.06	112.53	6.0	M_S, GS	6.2	Pocatello Valley (Ida.-Utah border)	15S, 18LS
43.	1975	Jun 30	18:54	44.69	110.62	6.1	M_L,GS	—	Yellowstone Park, Wyo.	19LS
44.	1976	Dec 08	14:40	44.76	110.80	5.5	m_b,GS	—	Yellowstone Park, Wyo., aftershock	7LS
45.	**1983**	**Oct 28**	**14:06**	**43.97**	**113.92**	7.3	M_S,GS	6.9	Borah Peak, Idaho	13LS, 21S
46.	1983	Oct 28	19:51	44.05	113.92	5.8	M_L	5.4	Borah Peak, Idaho, aftershock	13L, (15, 21)S
47.	1983	Oct 29	23:29	44.24	114.06	5.8	M_L	5.5	Borah Peak, Idaho, aftershock	13L, (15, 21)S
48.	1983	Oct 29	23:39	44.24	114.11	5.5	m_b,GS	—	Borah Peak, Idaho, aftershock	13LS
49.	1984	Aug 22	09:46	44.37	114.0[illegible]	5.8	M_L	[illegible]	Borah Peak, Idaho, aftershock	13L, (15, 22)S

Explanation:

The local time (Mountain Standard Time) for the earthquakes in this table is found by subtracting seven hours from Greenwich Mean Time. In some cases (War Time, Daylight Savings Time), the difference is six hours. Non-instrumental and instrumental earthquake locations are listed with one- and two-decimal-point accuracy, respectively. **Earthquakes accompanied by surface faulting have their origin time and location in bold print.**

Abbreviations for earthquake magnitude: M_{conv} = conventional magnitude, including M_{GR} (unified magnitude determined by Gutenberg and Richter), M_L (local magnitude), M_S (surface-wave magnitude), and m_b (body-wave magnitude); I/A = estimate of M_L based on intensity and/or felt area (values in parentheses based on authors' judgment here); REN = Reno, empirical estimate of M_{GR}; PAS = Pasadena, unspecified M_S or M_L (except 31 and 40, known to be M_L); BRK = Berkeley (all values here are M_L); GS = U.S. Geological Survey; M_w = moment magnitude (Hanks and Kanamori, 1979).

Key to references, including source of location (L) and size (S): 1. Williams and Tapper (1953); 2. Coffman and others (1982); 3. Arabasz and McKee (1979); 4. Townley and Allen (1939); 5. Doser (1989a); 6. Gutenberg and Richter (1954); 7. Engdahl and Rinehart (this volume); 8. Qamar and Stickney (1983); 9. Doser (1989b); 10. Jones (1975); 11. Doser (1990); 12. Qamar and Hawley (1979); 13. Dewey (1987); 14. Doser (1985); 15. Doser and Smith (1989); 16. Abe (1981); 17. Westaway and Smith (1989b); 18. Arabasz and others (1981); 19. Pitt and others (1979); 20. R. Uhrhammer (personal communication, 1990); 21. Richins and others (1987); 22. Zollweg and Richins (1985).

continuing interest (McCalpin and others, 1987; dePolo and others, 1989), especially in light of recent seismic waveform modeling by Doser (1989b) that indicates a main-shock focal mechanism with nearly pure strike slip, rather than normal slip. The waveform modeling implies a main-shock focal depth of 8 to 10 km, left-lateral slip on a plane striking N38° to 48°E, and a subsurface rupture length of about 11 km (Doser, 1989b). Strong aftershocks were recorded at regional distances (Table 2).

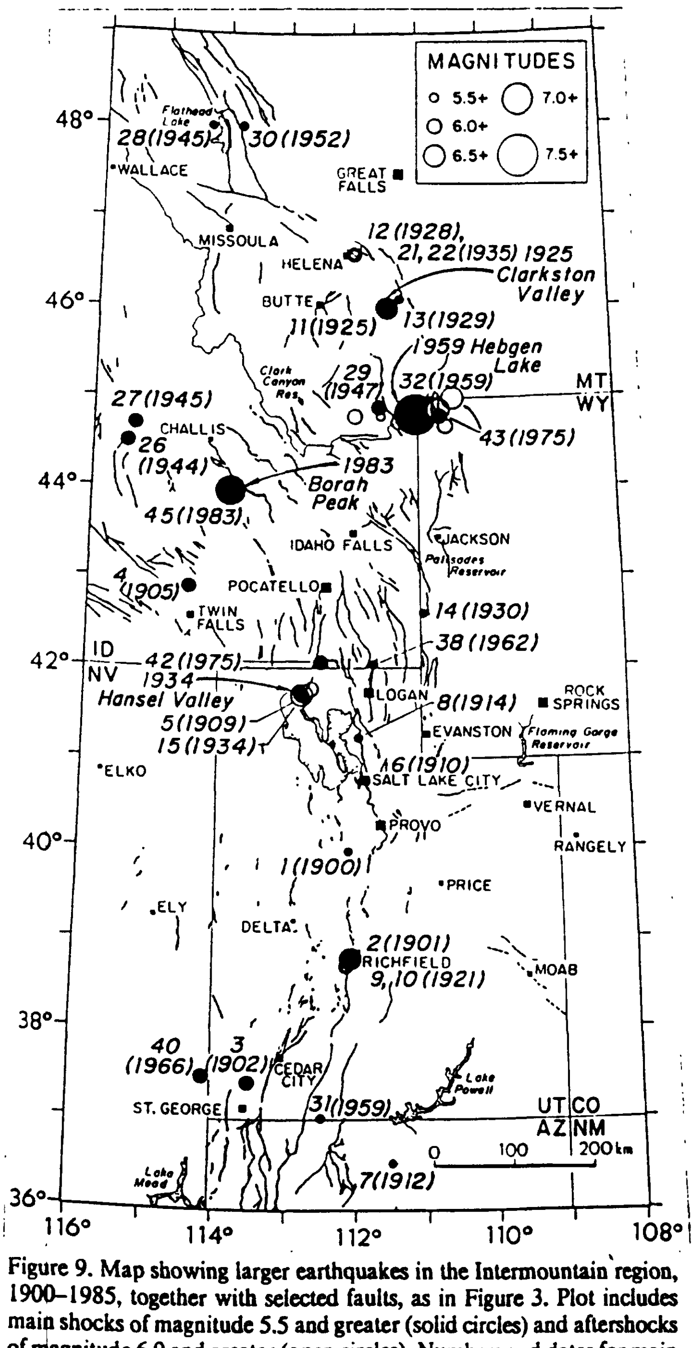

Figure 9. Map showing larger earthquakes in the Intermountain region, 1900–1985, together with selected faults, as in Figure 3. Plot includes main shocks of magnitude 5.5 and greater (solid circles) and aftershocks of magnitude 6.0 and greater (open circles). Numbers and dates for main shocks are keyed to Table 2. Four largest historical shocks known to exceed moment magnitude (M_W) 6.5 are indicated by name.

1959 Hebgen Lake, Montana, earthquake. The 1959 earthquake near Hebgen Lake in southwestern Montana (No. 32a, b, Table 2), directly west of Yellowstone National Park, was the first large normal-faulting earthquake in the Intermountain region in historical time and is distinguished as the largest recorded earthquake in the ISB (see special collection of papers in vol. 52, no. 1, of the *Bulletin of the Seismological Society of America,* 1962). The surface-wave magnitude (M_S) of 7.5 from Abe (1981) is a multistation estimate judged to be more accurate than the earthquake's previously estimated magnitude of 7.1, attributed to Pasadena by Tocher (1962). The occurrence of the M_S = 7.3 Borah Peak, Idaho, earthquake in 1983 (discussed below) prompted immediate comparison between the two large earthquakes. From comparisons of surface faulting (Hall and Sablock, 1985) and recorded seismograms (Bolt, 1984), the larger size of the Hebgen Lake earthquake was evident—confirmed by subsequent comparison of the two earthquakes' source parameters determined both seismically and geodetically (e.g., Barrientos and others, 1987). Restudy of Wood-Anderson seismograms at Berkeley and Pasadena for the 1959 earthquake and comparison with counterpart recordings for the 1983 earthquake led to a revised estimate of 7.7 to 7.8 for the local or Richter magnitude (M_L) of the 1959 earthquake and assignment of 7.2 (M_L) for the 1983 earthquake (Bolt, 1984). This relative size information is valuable. Given the large distances to Berkeley and Pasadena, however, the distance correlations—and hence the absolute values of M_L—are open to question.

The Hebgen Lake earthquake affected an area of 1.5 million km^2, reached MMI = X, caused 28 fatalities, and produced dramatic surficial geologic effects, including spectacular fault scarps, a catastrophic rockslide into the Madison River, basin subsidence of several meters, and hydrogeomorphic features associated with groundwater discharge (Witkind and others, 1962; Coffman and others, 1982). The earthquake produced a 26-km-long complex pattern of west- to northwest-trending normal faulting along the Hebgen and Red Canyon faults near the southern end of the Madison Range, where Laramide faults exert structural control (Witkind, 1964; Doser, 1985b). Trace lengths of surface faulting, including slip on the nearby north-south–trending Madison fault, sum to 61 km (Hall and Sablock, 1985). Maximum vertical displacement is variously cited as 6.7 m by dePolo and others (1989) and 5.5 ± 0.3 m by Bonilla and others (1984). Hall and Sablock (1985) use data from Witkind (1964) to estimate an average vertical displacement of 2.0 m on the Hebgen fault and 2.3 m on the Red Canyon fault.

Notable seismological details of the earthquake sequence (Doser, 1985b, 1989a) include: (1) the occurrence of the main shock as a multiple event, consisting of a shock of m_b = 6.3 at about 10 km depth followed 5 sec later by the principal M_S = 7.3 shock at 15 km depth; (2) nearly pure dip-slip motion during the main shock on one or more planes dipping 40° to 60°SW; and (3) the location of numerous strong aftershocks as large as M_S = 6.3 (Table 2) out to distances of 50 km from the main-shock epicenter. Barrientos and others (1987) recently reanalyzed the

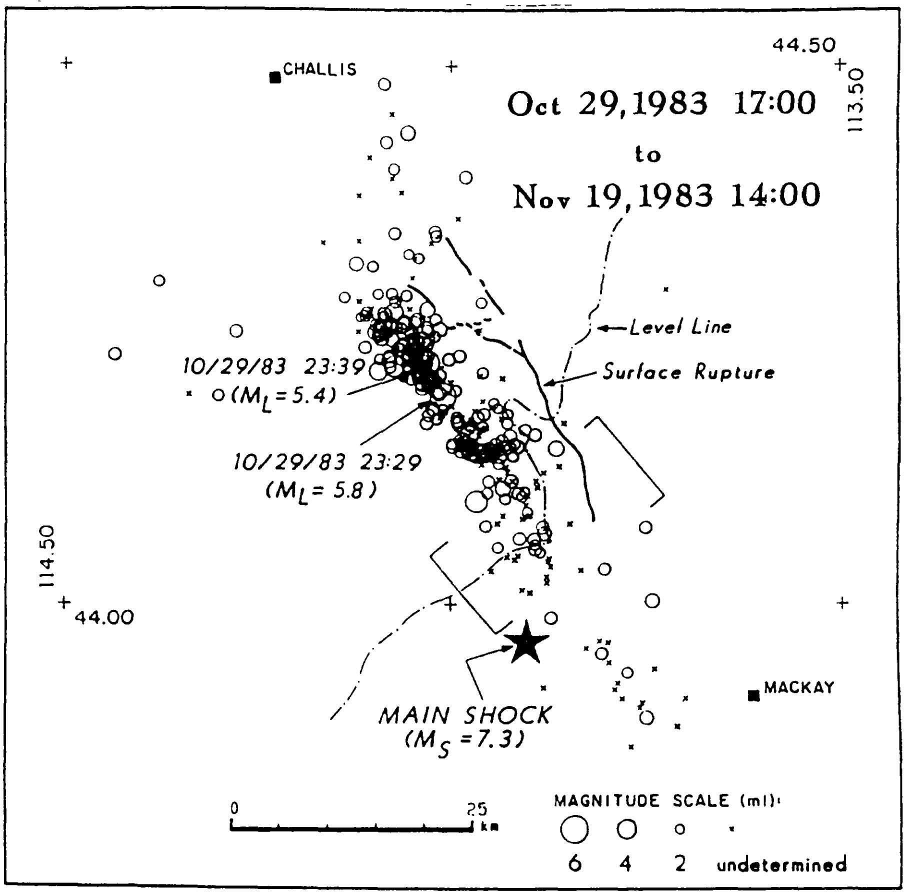

Figure 10. Map (from Richins and others, 1987) showing epicenters of the main shock and early aftershocks of the Borah Peak, Idaho, earthquake sequence, together with the trace of surface rupture. Brackets outline sample area for earthquakes shown in cross section of Figure 11 (middle); dot-dashed line, trace of geodetic level line from which data are displayed in Figure 11 (top).

static deformation field associated with the earthquake using a newly augmented geodetic data set. They interpret a complex source consisting of two en-echelon planes, 15 to 25 km long, which are coincident with the Hebgen and Red Canyon faults at the surface, extend to a depth of 10 to 15 km, dip 45° to 50°SW, and have coseismic dip slip of 7.0 and 7.8 m.

1983 Borah Peak, Idaho, earthquake. This 1983 surface-faulting earthquake (No. 45, Table 2) occurred in east-central Idaho, 60 km northwest of the Snake River Plain, in a sparsely settled area characterized by active late Quaternary basin-range faulting (Scott and others, 1985) but low historic seismicity (Dewey, 1987). The M_S = 7.3 earthquake is the second-largest historical earthquake in the Intermountain region. Importantly, the earthquake allowed abundant modern observations about the mechanics, subsurface rupture geometry, and seismic geology of a large normal-faulting earthquake (see special compendia of papers in U.S. Geological Survey Open-File Report 85-290, 1985, and vol. 77, no. 3, of the Bulletin of the Seismological Society of America, 1987).

The Borah Peak earthquake was felt over 670,000 km^2, reached MMI = VII at the nearby towns of Mackay and Challis (25 km and 65 km distant, respectively, from the main-shock epicenter), caused two deaths in Challis, and resulted in about $12.5 million of damage (Stover, 1985). The earthquake produced 36 km of surface faulting along the southwestern base of the Lost River Range, re-rupturing parts of the 140-km-long Lost River fault that had last broken about mid-Holocene time and a branch fault; vertical displacement along the new fault scarps reached a maximum of 2.7 m (0.8 m average), and net slip averaged 0.17 m of sinistral slip for every 1.00 m of dip slip (Crone and Machette, 1984; Crone and others, 1987). There is substantial information about the segmented behavior of the Lost River fault during the 1983 earthquake and about the fault's paleoseismology (see reviews by Crone and Haller, 1989, and dePolo and others, 1989).

Figure 10 (from Richins and others, 1987) illustrates the map distribution of aftershocks with respect to the main-shock epicenter and surface rupture. The earthquake sequence included sizable aftershocks (Table 2) but no foreshocks (Richins and others, 1987; Dewey, 1987). Seismic-waveform modeling by

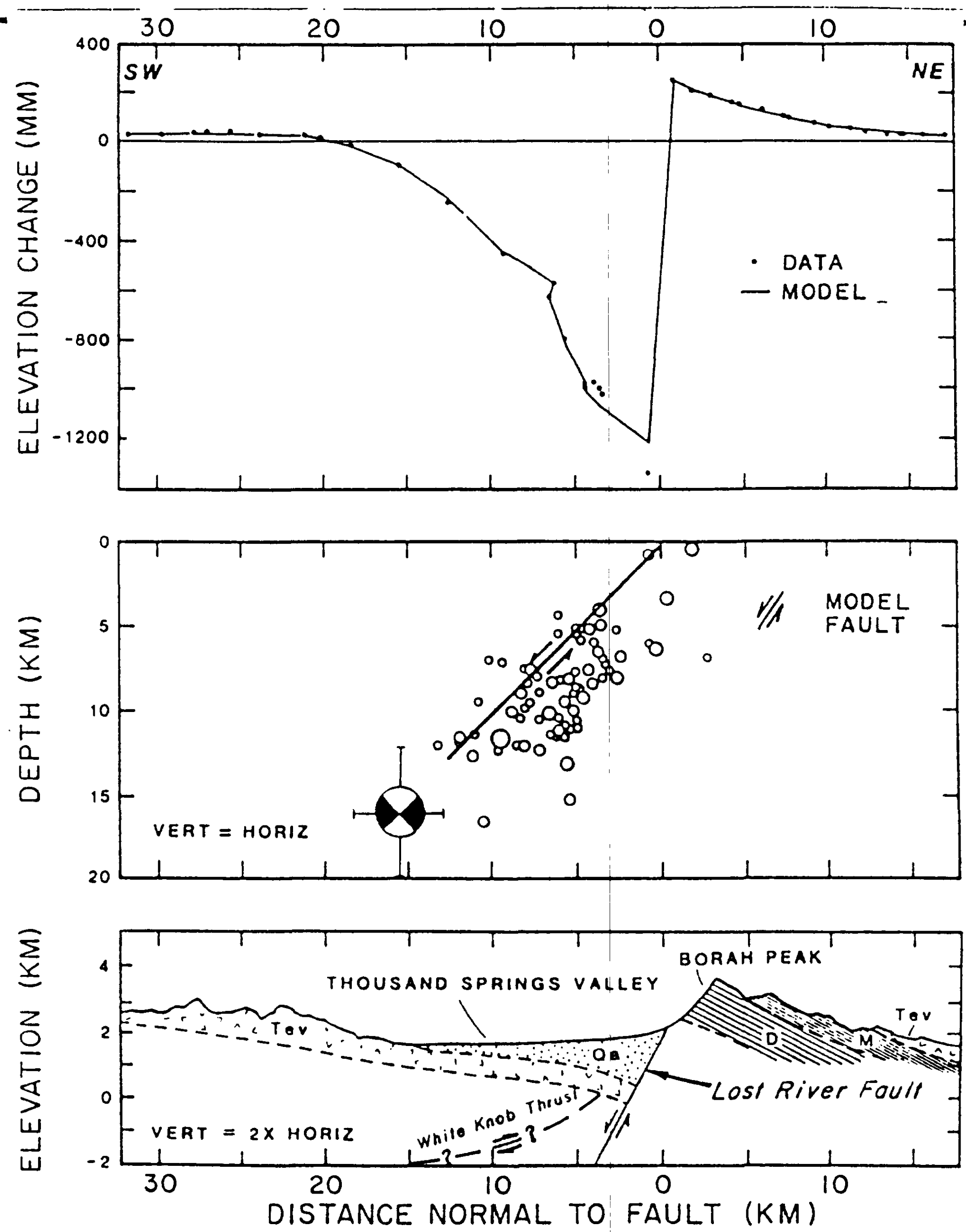

Figure 11. Diagram illustrating the subsurface geometry of faulting associated with the 1983 Borah Peak, Idaho, earthquake (after Stein and Barrientos, 1985). Top, plot of geodetically-observed coseismic elevation changes (dots) and the predicted elevation changes (line) of the coseismic dislocation model. Middle, cross section showing superposition of coseismic dislocation model, aftershock foci from the bracketed area shown in Figure 10, and the projection of the main-shock focus. Bottom, schematic geologic cross section from Bond (1978).

Doser and Smith (1985) indicates the main shock nucleated at a depth of about 16 km and propagated unilaterally northwestward toward the surface along a fault plane dipping 45° to 53° southwest (see Richins and others, 1987, Table 2, for a comparative tabulation of the main shock's source parameters). Figure 11, after Stein and Barrientos (1985), usefully illustrates some of the principal aspects of subsurface fault geometry and deformation. The data are displayed in transverse view to the Lost River fault (bottom) and are keyed to a northeast-southwest geodetic profile of coseismic elevation changes (top) along an irregular leveling route roughly transverse to the fault (Fig. 10). The middle panel shows a cross section of aftershock foci from Richins and others (1985), correlative with the bracketed sample area shown in Figure 10, together with the location and focal mechanism of the main shock and a planar dislocation model that matches the observed surface deformation. Focal mechanisms for 47 aftershocks suggest that most of the aftershock foci reflect complex fracturing on secondary structures adjacent to the main fault plane rather than seismic afterslip on a simple main-shock rupture plane (Richins and others, 1987).

Barrientos and others (1987) have refined the dislocation model represented in Figure 11 using supplementary geodetic data, but changes to the illustration would be slight. Their preferred model has a planar fault in the depicted section that dips 49° (instead of 47° as shown), extending to a depth of 14 km; the northern part of the fault is modeled with a dislocation extending only to 6 km depth. The modeling yields an average dip-slip displacement of 2.1 m on the southern dislocation and 1.4 m on the northern one. An important point is that the geodetic data do not permit a listric fault geometry (see also Stein and Barrientos, 1985). Source parameters determined by Barrientos and others (1987) from this geodetic modeling (slip = 1.4 to 2.1 m, static stress drop = 30 bars, moment = 2.9×10^{26} dyne-cm) are consistent with seismically-determined values for the earthquake (e.g., Doser and Smith, 1985: slip = 1.4 m, static stress drop = 17 bars, moment = 2.1×10^{26} dyne-cm).

Other significant earthquakes

Table 2 and Figure 9 give a succinct overview of other significant earthquakes in the Intermountain region besides the four largest just described. Using Table 2 for reference, to minimize repetition of information, we complete this section by mentioning the other main shocks of magnitude 6 and some notable smaller shocks.

Southern ISB. The 1901 Southern Utah (Richfield) earthquake (Event No. 2) appears to be the second largest historical shock in the southern ISB, although its equivalent magnitude and precise epicenter are not well known. The earthquake reached MMI = IX, was felt over 130,000 km^2, caused substantial damage at several towns and produced ground cracks (but no documented surface faulting), local liquefaction, and extensive rockslides (Williams and Tapper, 1953). In late 1921, following two and a half weeks of foreshock activity, the Elsinore area, 10 km southwest of Richfield, was struck by two damaging earthquakes of about magnitude 6¼ (Events No. 9 and 10), separated by 50 hours and with an intervening shock of magnitude 5¾ (Pack, 1921; Arabasz and Julander, 1986). Southwestern Utah was significantly affected in 1902 by a damaging earthquake (MMI = VIII) centered in Pine Valley (Event No. 3) and in 1966 by a sizable earthquake (m_b, USGS = 6.1; M_L, Univ. of Utah = 5.6) close to the Nevada-Utah border (Event No. 40) (Arabasz and others, 1979; Coffman and others, 1982). The 1966 event was notable for its strike-slip focal mechanism in a basin-range setting (Smith and Sbar, 1974; Rogers and others, this volume).

Two other magnitude 6 shocks in the southern ISB occurred in the Utah-Idaho border area. This includes a strong (MMI = IX) earthquake in 1909 (Event No. 5), assumed to have originated in the Hansel Valley area (Williams and Tapper, 1953), and the damaging M_L = 6.0 Pocatello Valley (Idaho-Utah border area) earthquake of 1975 (Event No. 42), notable for its discordant relation with the surface geology (Arabasz and others, 1981) and its space-time pattern of precursory seismicity (Arabasz and Smith, 1981). Another noteworthy earthquake is the M_L = 5.7 Cache Valley (Logan) earthquake of 1962 (Event No. 38), the most damaging yet in Utah's history (Cook, 1972; Rogers and others, 1976; Westaway and Smith, 1989b) and the only sizable earthquake in the Utah region for which good strong-motion recordings—one three-component set—currently exist (Smith and Lehman, 1979; Westaway and Smith, 1989a, b).

Central ISB. Four magnitude 6 main shocks have occurred in the Central ISB. Shocks of magnitude 6.1 (Event No. 26) and magnitude 6.0 (Event No. 27) occurred seven months apart in 1944 and 1945 in central Idaho along the eastern flank of the Idaho batholith, causing only minor damage in the remote mountainous setting (Coffman and others, 1982). Revised instrumental locations place both earthquakes relatively close to each other, suggesting a possible main-shock/large-aftershock relation (Dewey, 1987). The magnitude 6¼ Virginia City, Montana, earthquake of 1947 (Event No. 29) caused considerable damage in the Madison Valley (Coffman and others, 1982). The earthquake occurred about 50 km west-northwest of the 1959 Hebgen Lake earthquake and has been studied by Doser (1989a). Large aftershocks of the 1959 earthquake extended into the Yellowstone National Park region (Fig. 9) where an independent main shock occurred later in 1975. The M_L = 6.1 Yellowstone Park earthquake of 1975 (Event No. 43) caused moderate disruption in the national park (Coffman and others, 1982) and provided valuable information on the seismotectonics of the Yellowstone caldera (Pitt and others, 1979). A strong earthquake (MMI = VII), probably in the magnitude 5 range, which caused damage at Shoshone, Idaho, in 1905 (Event No. 4) (Coffman and others, 1982), has an uncertain epicenter but is significant for its possible association with the relatively aseismic Snake River Plain.

Northern ISB. The 1935–1936 Helena earthquakes (Event nos. 20 to 25) were part of a vigorous swarm of more than two thousand felt earthquakes that occurred within 10 to 25 km of Helena, between October 1935 and December 1936, within a poorly defined structural zone extending from Helena northwestward toward Marysville, an area of anomalously high heat flow (Smith and Sbar, 1974; Doser, 1989a). Earthquakes of magnitude 6¼ (Event No. 21) and magnitude 6 (Event No. 22) in October 1935 caused four deaths and severe damage in Helena (Coffman and others, 1982). Some important strong-ground-motion records were recorded locally (Westaway and Smith, 1989a). Other earthquakes in the northern ISB listed in Table 2, which caused only minor damage (Coffman and others, 1982; Qamar and Stickney, 1983), include: a 1928 shock near Helena (Event No. 12), perhaps a precursor to the 1935 swarm; a 1929 shock near Lombard (Event No. 13), 9 km north of Clarkston, possibly a late aftershock of the 1925 Clarkston Valley earthquake; and two shocks in the Flathead Lake region in 1945 (Event No. 28) and 1952 (Event No. 30).

DETAILED SEISMICITY

In earlier sections we gave an overview of the regional-scale patterns of earthquake activity in the ISB (Fig. 1) and described the largest earthquakes that have occurred historically (Fig. 9). In this section, we describe finer details of the spatial distribution of the seismicity portrayed in Figure 1, outlining what is known about the association of the seismicity with geologic structure. Observational data basically come from either long-term monitoring with telemetered seismic networks (Fig. 8B) or focused short-term monitoring using temporary arrays of portable seismographs. Some notable characteristics of observed seismicity throughout the ISB are: (1) the diffuse epicentral scattering of background earthquakes, with weak correlation to major active faults; (2) the conspicuous seismic quiescence of many major faults or fault segments that have been active in Holocene and late Quaternary time; (3) the predominance of focal depths shallower than 20 km; and (4) the prevalence of normal and oblique-normal seismic slip, but with local strike slip and reverse slip.

Southern ISB

Detailed summaries of the instrumental seismicity for major parts of the southern ISB have recently been given by Arabasz and others (1987) for the Wasatch Front area, by Arabasz and Julander (1986) for the Basin and Range–Colorado Plateau transition in central Utah, and by Wong and Humphrey (1989) for the Colorado Plateau. Here, we selectively adapt from and add to those summaries.

The recorded seismic history of the southern ISB has been distinctively characterized by abundant small- to moderate-sized earthquakes (magnitude $\leqslant$ 6.6) without a truly large surface-faulting earthquake, despite the widespread presence of late Pleistocene and Holocene fault scarps. The single instance of historical surface faulting in the southern ISB at Hansel Valley in 1934 (discussed earlier) was for an earthquake only slightly above the threshold of surface faulting. Thus, there is a lack of instrumental information relating to large-scale seismic slip on major faults in this region. Available earthquake observations may chiefly reflect seismic deformation on secondary structures.

The seismicity of the Wasatch Front area shown in Figure 12 displays no simple correlation between the distribution of background earthquakes and the traces of the numerous active faults, except for the general parallelism of this part of the ISB with the Wasatch fault. The depth above which 90 percent of the well-located earthquakes lie varies locally from about 11 to 17 km (Arabasz and others, 1987). Well-located shocks of magnitude 2.0 and greater in this area from 1962 to 1986 have a distinct peak in their depth distribution between 4.5 and 7.5 km (Bjarnason and Pechmann, 1989). An anomalously deep earthquake of M_L = 3.8 occurred at a depth of 90 km beneath northern Utah in 1979 (see Wong and Chapman, 1990).

The seismicity pattern of Figure 12B is representative of instrumental seismicity in the area since 1962 (see Arabasz and others, 1980); spatial clustering is due more to cumulative stationary activity than to isolated temporal bursts. An inverted Y-shaped pattern of clustered earthquakes on the Idaho-Utah border west of the Wasatch fault began to develop several months after the 1975 M_L = 6.0 Pocatello Valley earthquake (No. 42, Fig. 9) and persists to the present. Early aftershocks of the 1975 earthquake occurred mostly north of the Idaho-Utah border (Arabasz and others, 1981). Special studies of post-1975 earthquakes within the inverted Y-pattern (Jones, 1987; Chen, 1988) show that well-located foci are mostly shallower than 8 to 12 km deep, scatter beneath both horsts and grabens of the local basin-range structure, and display diverse seismic slip.

A linear north-south belt of seismicity about 15 to 40 km east of and parallel to the Wasatch fault (Fig. 12) is poorly understood but may be mechanically related to crustal flexure associated with the Wasatch fault and involving non-elastic mechanisms (Zandt and Owens, 1980; Owens, 1983). This seismicity is known to coincide in map view with the eastern leading edges of several Laramide thrust sheets (Smith and Bruhn, 1984). In cross-section view, however, the earthquake foci are diffusely scattered to about 20 km depth (Arabasz and others, 1987; Owens, 1983). In the northern part of the belt, the earthquakes lie east of the west-dipping East Cache and Wasatch faults, perhaps partly on a synthetic fault (Westaway and Smith, 1989b); south of 41°20′N, the epicentral belt follows a series of small, late Cenozoic structural basins within the Middle Rocky Mountains (Sullivan and others, 1988). In the lower right part of Figure 12, the prominent arcuate pattern of seismicity east of the Wasatch fault is mining related (see Induced seismicity).

What about earthquakes along the Wasatch fault itself? In addition to having no historic surface rupture, despite recurrent late Pleistocene and Holocene surface faulting on multiple segments (Machette and others, 1989; Schwartz and Coppersmith, 1984), the Wasatch fault has had little historical seismicity. As many as two, and perhaps no, earthquakes as large as magnitude 5 have occurred on the Wasatch fault in historical time (Arabasz and others, 1987). The most recent surface rupture occurred about 400 years ago (Machette and others, 1989) on a 40-km-long segment immediately north of Nephi (Fig. 12A). In terms of contemporary seismicity, Figure 12B shows a remarkable paucity of microseismicity along most of the Wasatch fault. There are a few local clusters of epicenters along the fault north of Brigham City, and more prominent clusters just west of the fault in the vicinity of Salt Lake City, at the northern end of Utah Valley (~40°20′N), in the vicinity of Goshen Valley (~40°00′N), and in a broadly scattered zone at the southern end of the Wasatch fault. For Goshen Valley and the southern Wasatch fault, hypocenters and corresponding focal mechanisms from portable-array studies show that background earthquakes west of the fault are not occurring on either a listric or a simple planar projection of the fault (Arabasz and Julander, 1986). Elsewhere along the Wasatch fault, cross sections suggest that very few well-located foci could be interpreted to lie on the fault—if one believes that the fault is a planar structure of moderate dip (Arabasz and others,

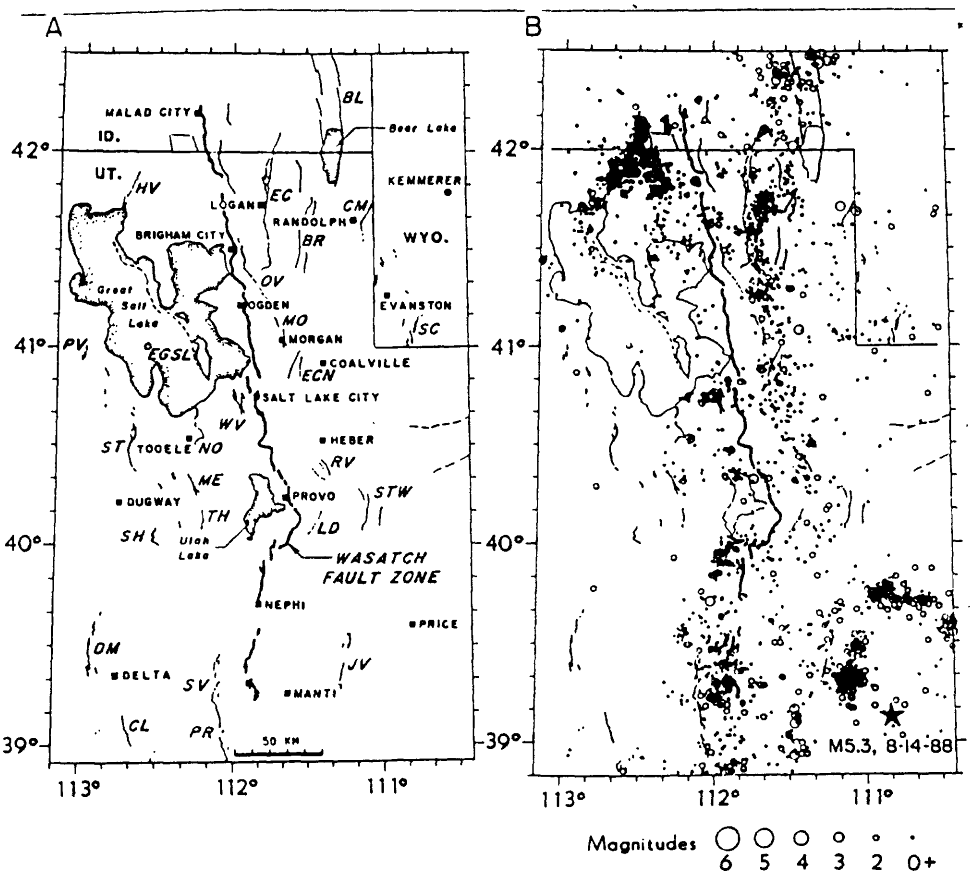

Figure 12. Active faulting and seismicity in the Wasatch Front area, outlined in Figure 3, from Arabasz and others (1987). A, Map showing traces of late Quaternary faulting, abbreviated as follows: BL = Bear Lake; BR = Bear River Range; CL = Clear Lake; CM = Crawford Mts.; DM = Drum Mts.; EC = East Cache; ECN = East Canyon; EGSL = East Great Salt Lake; HV = Hansel Valley; JV = Joes Valley; LD = Little Diamond Creek; ME = Mercur; MO = Morgan; NO = Northern Oquirrh; OV = Ogden Valley; PR = Pavant Range; PV = Puddle Valley; RV = Round Valley; SC = Sulphur Creek; SH = Sheeprock Mts.; ST = Stansbury Mts.; STW = Strawberry Valley; SV = Scipio Valley; TH = Topliff Hill; WV = West Valley. B, Epicenter map of all earthquakes located by the University of Utah Seismograph Stations in the Wasatch Front area, July 1, 1978 to December 31, 1986; star indicates location of 1988 San Rafael Swell, Utah, main shock described in Figure 14.

1987). But the same data are admittedly either inadequate or ambiguous (e.g., Pechmann and Thorbjarnardottir, 1984) for interpreting subsurface association with a listric projection of the fault.

Portable-seismograph studies throughout the transition zone between the Basin and Range and Colorado Plateau provinces in central and southwestern Utah reveal that low-angle structural discontinuities appear to play a fundamental role in separating locally intense upper-crustal seismicity above 6 to 8 km depth from less frequent background earthquakes at greater depth (Arabasz and Julander, 1986). This is illustrated in Figure 13A, where spatially discontinuous seismicity with depth beneath the Sevier Valley near Richfield, Utah, coincides with a low-angle detachment inferred from seismic-reflection data. The earthquakes beneath the northwestern side of the valley are background events recorded in 1981; those beneath the southeastern side are aftershocks of an M_L = 4.0 earthquake in May 1982. Companion results from a portable-seismograph study of an earthquake swarm sequence ($M_L \leq 4.7$) in October 1982 near Soda Springs, Idaho (Fig. 13B), similarly show a depth distribution of upper-crustal earthquakes apparently influenced by preexisting low-angle structures. Instead of being associated, as expected, with late Cenozoic basin-range faulting along the active Bear Lake fault, the seismicity is associated with secondary faults within a northwest-trending near-vertical zone in the hanging-wall block. Marked changes in the vertical distribution of foci coincide with pre-Neogene thrust faults. Focal mechanisms sampled from both above and below the Meade thrust indicate a predominance of strike slip on northwest-trending, steeply-dipping fault planes, with no evidence for seismic slip on a low-angle plane (see Arabasz and Julander, 1986).

East of the southern ISB, scattered seismicity within the Colorado Plateau (Figs. 1 and 2), described by Wong and Humphrey (1989), occasionally reaches the magnitude 6 range

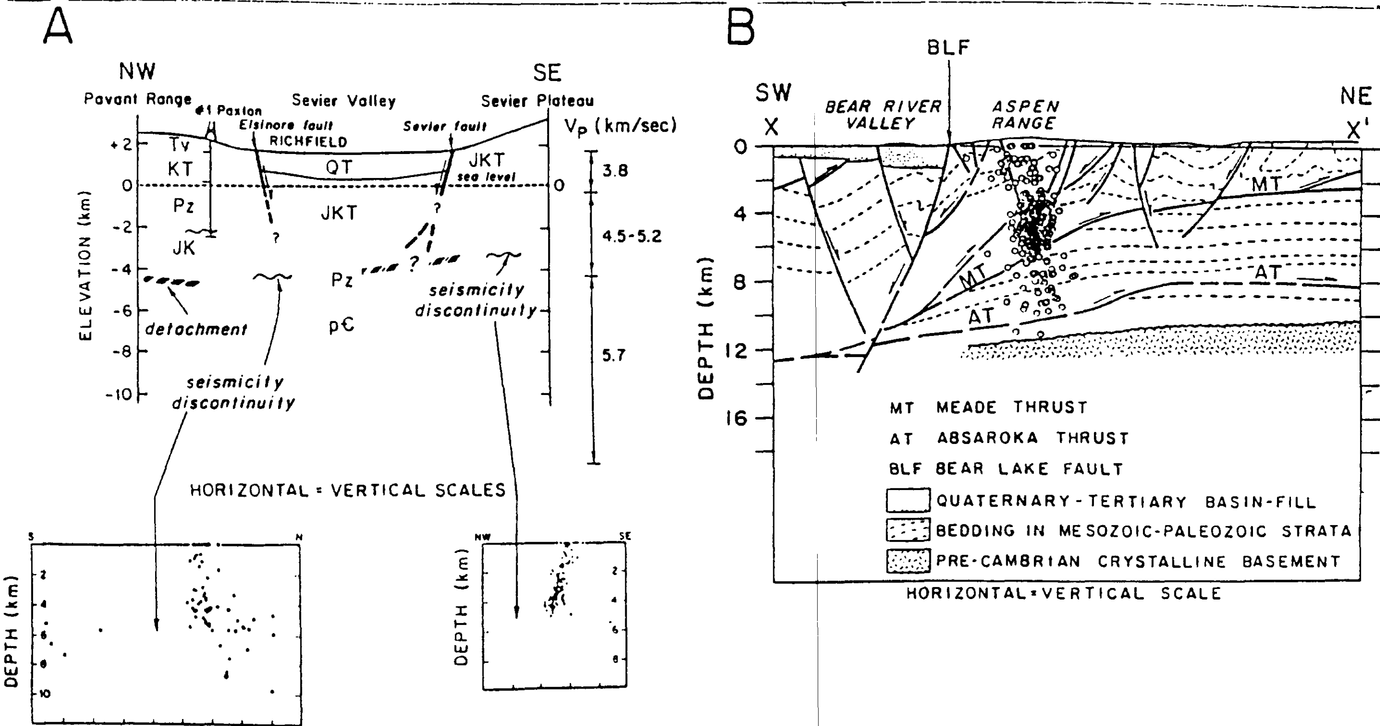

Figure 13. Results of two selected earthquake field studies carried out by the University of Utah illustrating the influence of low-angle structural discontinuities on seismicity, from Arabasz and Julander (1986). A, Schematic geologic cross section across the Sevier Valley near Richfield, Utah, showing spatially discontinuous seismicity with depth coincident with the location of a low-angle detachment inferred from seismic-reflection data. V_P = P-wave velocity; T_V = Tertiary volcanic rocks; other abbreviations, standard for geologic ages of rocks. B, Cross section showing the association of swarm seismicity in late 1982 near Soda Springs, Idaho, with geologic structure; earthquake data from Richins and others (1983) are superposed on a generalized geologic cross section from Dixon (1982) based on seismic-reflection profiling.

and appears chiefly to reflect normal to lateral seismic slip on buried Precambrian basement faults without evident surface expression. Focal depths predominate above 15-20 km, but the Colorado Plateau is distinctive in having observed seismicity in the lower crust, and locally 40 to 60 km deep in the uppermost mantle (Wong and Humphrey, 1989; Wong and Chapman, 1990). Figure 14 illustrates details of what may be a typical, moderate-sized crustal earthquake within the Colorado Plateau. The August 1988 San Rafael Swell earthquake (M_L = 5.3) involved oblique-normal slip on a buried Precambrian basement fault in an area of minimal historical seismicity where there are no active faults mapped in the overlying 3-km-thick sedimentary cover rocks of Mesozoic and Paleozoic age (Nava and others, 1988). Figure 14 also usefully illustrates some prerequisites for associating seismicity with geologic structure: (1) local seismographic control for hypocentral resolution, especially for precise focal depths; (2) sufficient seismicity for defining the spatial geometry of one or more active structures; and (3) a reliable focal mechanism for correlating with the geometry and sense of slip inferred from the seismicity.

Noteworthy microearthquake studies in the southern Utah–northern Arizona area have been reported by Johnson and Sbar (1987) and Kruger-Knuepfer and others (1985). Both studies present valuable focal-mechanism information relevant to regional stress orientations. Other key studies that summarize and discuss significant focal-mechanism information for the southern ISB include those by Bjarnason and Pechmann (1989), Wong and Humphrey (1989), Arabasz and Julander (1986), Zoback (1983), Arabasz and others (1980), Smith and Lindh (1978), and Smith and Sbar (1974). Besides implications for stress state (discussed earlier), the focal mechanisms also provide information on fault kinematics. Seismic slip predominates on fault segments of moderate (>30°) to steep dip. There is yet no convincing evidence, in the form of clustered earthquake foci and corroborating focal mechanisms, for seismic slip on either a downward-flattening or a low-angle normal fault in this region.

Central ISB

The central part of the ISB is distinctive in two regards. First, it has had two large surface-faulting earthquakes in historical time—the 1959 Hebgen Lake and 1983 Borah Peak earthquakes. Second, its seismicity reflects the apparent influence of the Y-SRP volcanic system. In map view this is illustrated by the arcuate pattern of seismicity described in the Introduction and shown in detail in Figure 15. This arcuate, parabolic pattern of earthquakes has been hypothesized to reflect lateral changes in deviatoric stresses in the wake of the relative passage of the Yellowstone hot spot (discussed in detail below), now centered beneath the Yellowstone caldera (Smith and others, 1985, in preparation; Anders and others, 1989; Blackwell, 1989). The seismogenic potential of the lithosphere axial to the SRP appears to be affected out to distances of more than 100 km. In Figure 15, for example, note the relative aseismicity of the SRP and the increase in seismicity away from the SRP along the trend of faults transverse to the plain.

Using Figure 15 as a guide, let us consider a counterclockwise circuit of the central ISB. Seismicity in the Utah-Idaho border region (already discussed) is near our defined juncture between the southern and central parts of the ISB. North of the Utah border there is an evident discordance between the northeast-trending seismicity belt and the northwest-trending late Cenozoic normal faulting. A refined compilation of earthquake locations verifies the regional seismicity pattern (Richins and Arabasz, 1985). On a local scale, the example of Figure 13B, from near Soda Springs in southeastern Idaho, illustrates the occurrence of small to moderate-sized background earthquakes near—but not on—one of the major active faults in this region. Portable-seismograph studies by the U.S. Bureau of Reclamation, in 1982–1983, of the region between Soda Springs and Jackson showed no obvious correlation between scattered microseismicity and local surficial faulting (Piety and others, 1986). Well-located earthquakes ($M_L \leqslant 3.0$) extended to 16 km depth, exhibited mostly normal slip, were apparently unaffected in their depth distribution by the presence of old thrustbelt structure, occurred abundantly within Precambrian basement below about 5 to 10 km depth, and appeared to be associated, in part, with buried normal faults having no surface expression (Piety and others, 1986).

Seismicity of the Idaho-Wyoming border area south of Yellowstone Park, encompassing the so-called Teton–Jackson Hole–southern Yellowstone region, has been summarized by Doser and Smith (1983) and by Smith and others (1990a, b) and is the target of network monitoring by the U.S. Bureau of Reclamation's Jackson Lake network (JL, Fig. 8B). Results described both by Doser and Smith (1983) from regional monitoring and

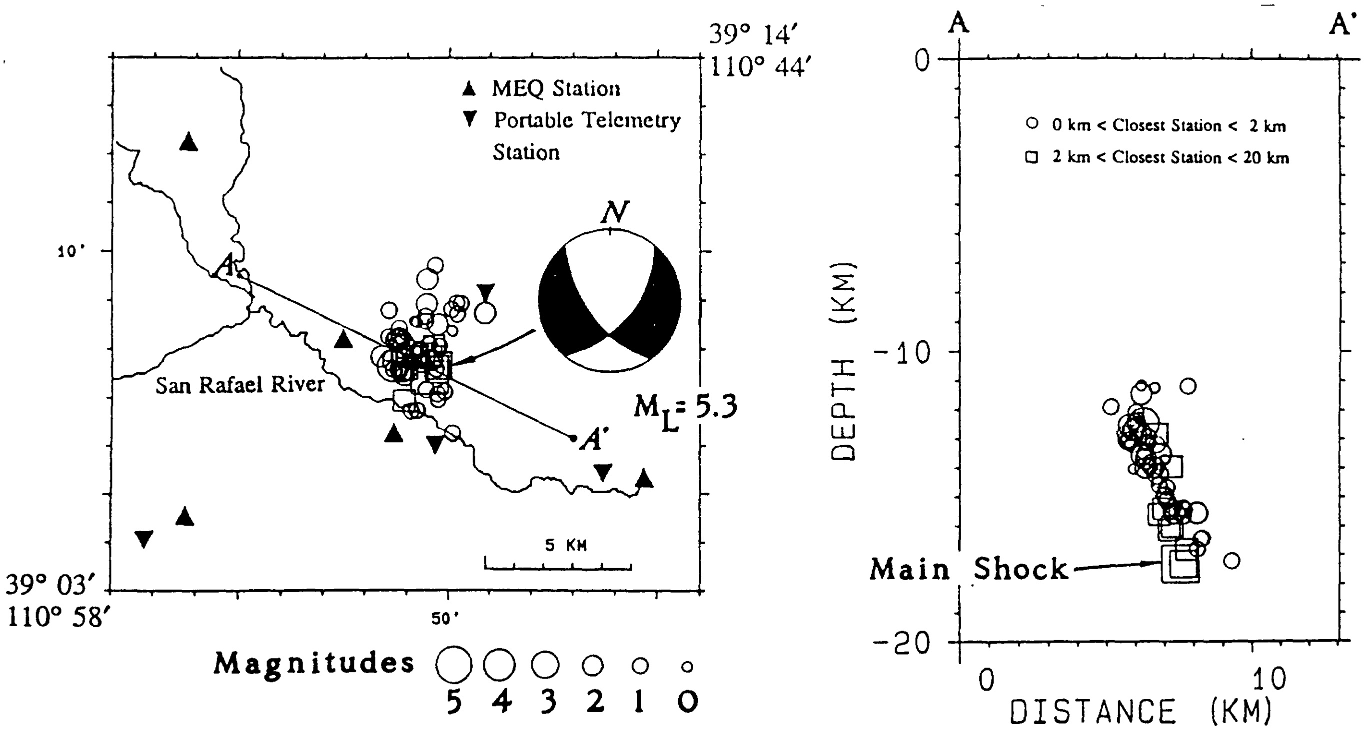

Figure 14. Details of the San Rafael Swell, Utah, earthquake sequence, August 14, 1988 to March 31, 1989, whose location is shown in Figure 12, adapted from Nava and others (1988) using revised, unpublished data from J. C. Pechmann and S. J. Nava, University of Utah. Left, map showing epicenters of the 66 best-located earthquakes in the sequence together with the main-shock focal mechanism (lower-hemisphere, compressional quadrant black). Right, cross section of earthquake foci projected onto the plane of line A-A′ (left), the projection for which planar clustering was best defined; the earthquakes define a zone dipping 60° ± 5° SE, with a downdip extent of 6 km. The main-shock focal mechanism has a corresponding nodal plane striking N39°E and dipping 62°SE, with a slip-vector rake of 29° (down to the NE); uncertainties for that plane range from N20°E to N42°E in strike, 44° to 80° in dip to the SE, and 21° to 59° in slip-vector rake.

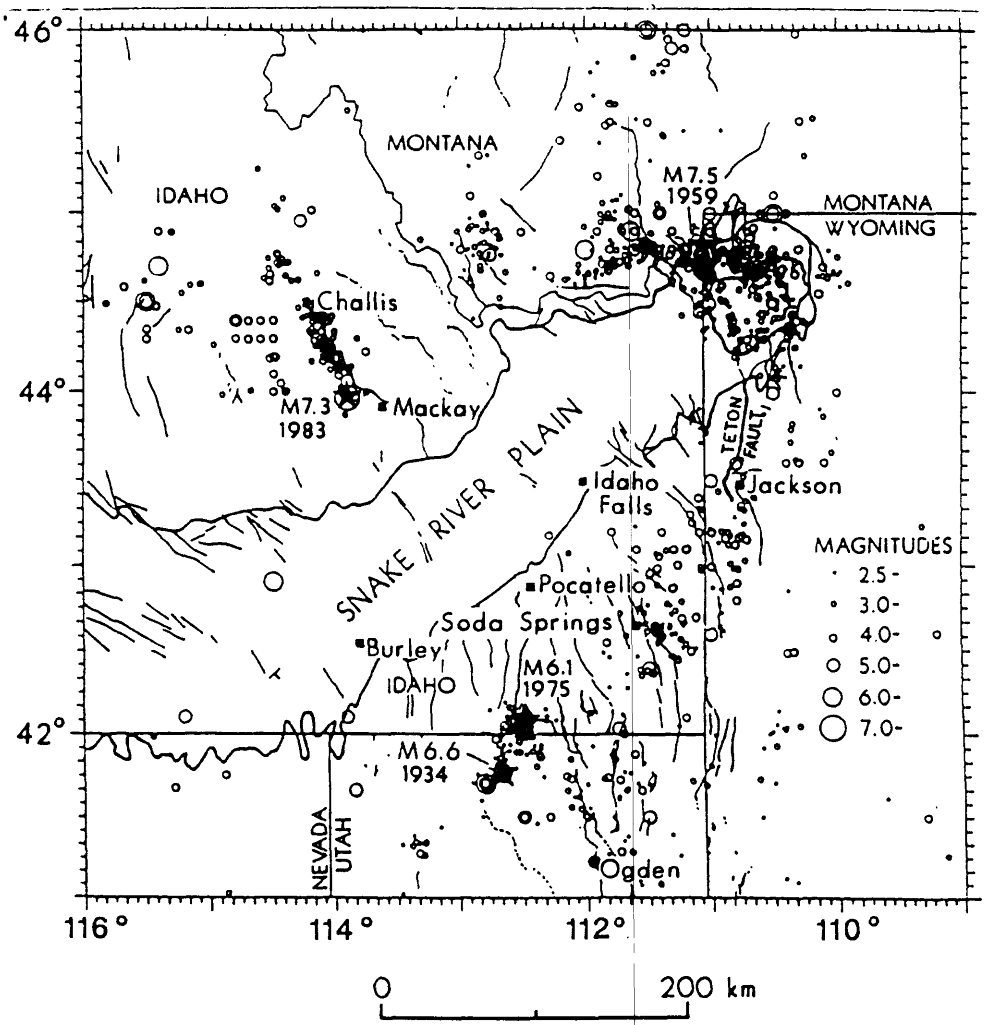

Figure 15. Map showing seismicity, 1900 to 1985, and selected Cenozoic faults of the central Intermountain seismic belt, from Smith and others (in preparation). Larger historical main shocks, as in Figure 9, shown for reference. Note the arcuate, parabolic pattern of seismicity flanking the Yellowstone–Snake River Plain basalt-rhyolite volcanic province.

local microearthquake studies and by Wood (1988) from the Jackson Lake network consistently depict: (1) diffusely scattered seismicity that correlates poorly with late Cenozoic normal faulting and, in an uncertain way, with relict thrustbelt structure in the subsurface; (2) seismic quiescence of the central part of the Teton fault, which had the greatest prehistoric surface displacements along the fault; (3) focal depths shallower than 15 km for most of the local earthquakes; and (4) fault-plane solutions with normal to strike slip reflecting general east-west extension. On a regional scale, perhaps the most noteworthy feature of the Teton region is its appearance as a distinct seismic gap in the ISB (Fig. 1), accentuated by the presence of the large and active Teton fault, which has had up to 50 m of late Quaternary displacement (Smith and others, 1990a, b).

Intense swarms of shallow earthquakes and occasional moderate-sized earthquakes as large as the $M_L = 6.1$ earthquake in 1975 near Norris Junction in Yellowstone Park (No. 43, Table 2) characterize the seismicity of the Yellowstone National Park region (Smith and others, 1977; Pitt, 1987; Nagy and Smith, 1988). The Yellowstone region is the site of one of the world's largest and most active hydrothermal-volcanic systems. During the past two million years, three catastrophic volcanic eruptions have expelled more than 3,500 km^3 of rhyolitic ashflow tuffs forming three calderas, and another 3,000 km^3 of similar material were extruded between explosive eruptions (Christiansen, 1984). A preponderance of geophysical evidence suggests that the seismicity of the Yellowstone region is directly influenced by the presence of magmas, partial melts, and hydrothermal activity at mid- to upper-crustal depths (Smith and others, 1974, 1977; Smith and Braile, 1984). Seismic slip on the boundaries of small upper-crustal blocks may reflect a combination of deformation caused by local transport of magma and hydrothermal fluid and by the regional tectonic stress field.

Figure 16 shows a representative view of background seismicity in the Yellowstone–Hebgen Lake region, together with an outline of the youngest, 600,000-year-old caldera. Earthquake clusters extend eastward from Hebgen Lake, Montana, along an east-west trend into Yellowstone National Park where they take on a northwest trend along distinct seismic zones about 25 km long that cross the caldera boundary. Within the caldera, earthquakes have not exceeded magnitude (M_L) 5.0 and generally have scattered epicenters; in the western part of the caldera, northwest-trending clusters of epicenters, together with aligned volcanic vents, may be related to buried, but still active, Quaternary faults (Christiansen, 1984). In several cases, there are good correlations between earthquake swarms and major changes in hydrothermal activity (Pitt and Hutchinson, 1982). Local faulting along the west side of Yellowstone Lake has Holocene displacements and appears to be seismically active. South of this area, seismicity has a general north-south trend as it extends southward into the Teton region. Older basin-range structure is inferred to have influenced the Quaternary tectonics of the Yellowstone region. Parts of the Gallatin and Teton normal fault systems, which generally have a northerly trend outside the Yellowstone region, presumably lie beneath the area now covered by the Quaternary volcanics of the Yellowstone Plateau.

Focal depths show conspicuous variations across the Yellowstone caldera (Fig. 17). Maximum focal depths outside the caldera are generally less than 15 to 20 km, and mostly less than 5 km beneath the inner caldera (Smith and others, 1977; Nagy and Smith, 1988). This pattern of earthquake shallowing suggests a thin layer of seismogenic brittle upper crust beneath the thermally active inner caldera. Rheologic considerations (e.g., Smith and Bruhn, 1984) imply that below about 5 km, the crust is in a quasi-plastic ductile state at temperatures in excess of 350°C, incapable of supporting large stresses. Note that the M_L = 6.1 earthquake in 1975 occurred along the caldera's northwest boundary.

Continuing our circuit of the central ISB in Figure 15, we next move westward from the Yellowstone region—following an

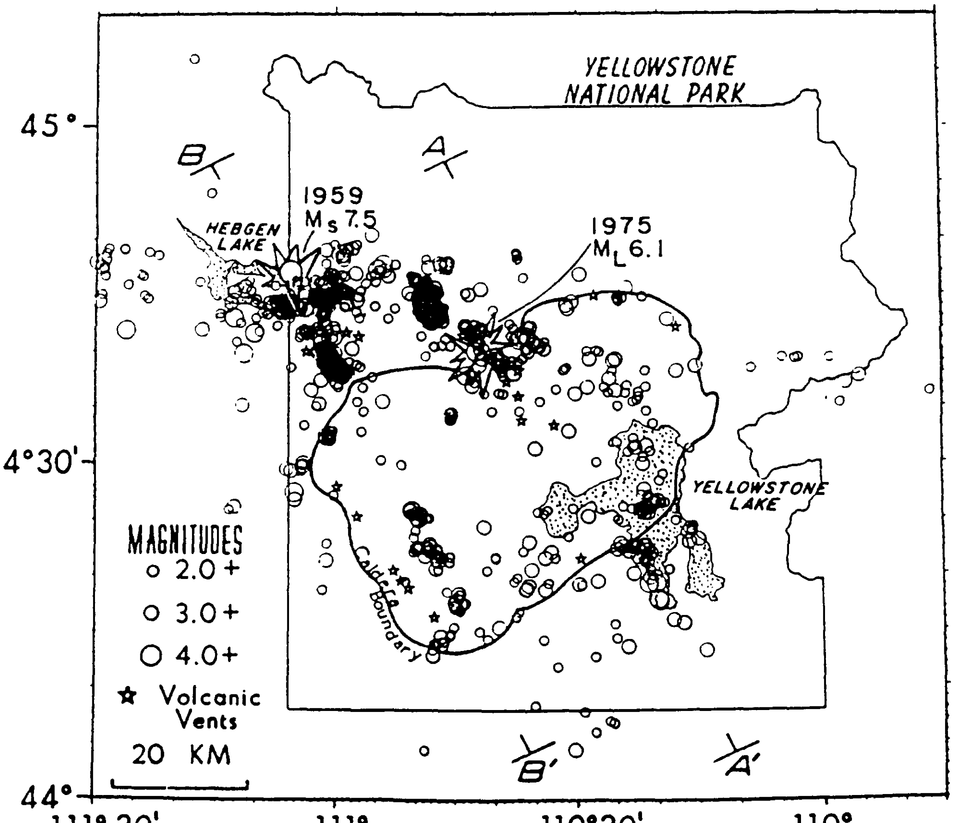

Figure 16. Map showing representative seismicity in the Yellowstone region of magnitude (M_L) 2.0 or greater, from compilations by the U.S. Geological Survey for 1973 to 1981 (Pitt, 1987) and by the University of Utah for 1984 to 1989 (e.g., Peyton and Smith, 1990). A-A′ and B-B′ define the locations of cross sections shown in Figure 17. Boundary of the Yellowstone caldera and epicenters of 1959 and 1975 main shocks (starbursts), as in Figure 9, shown for reference.

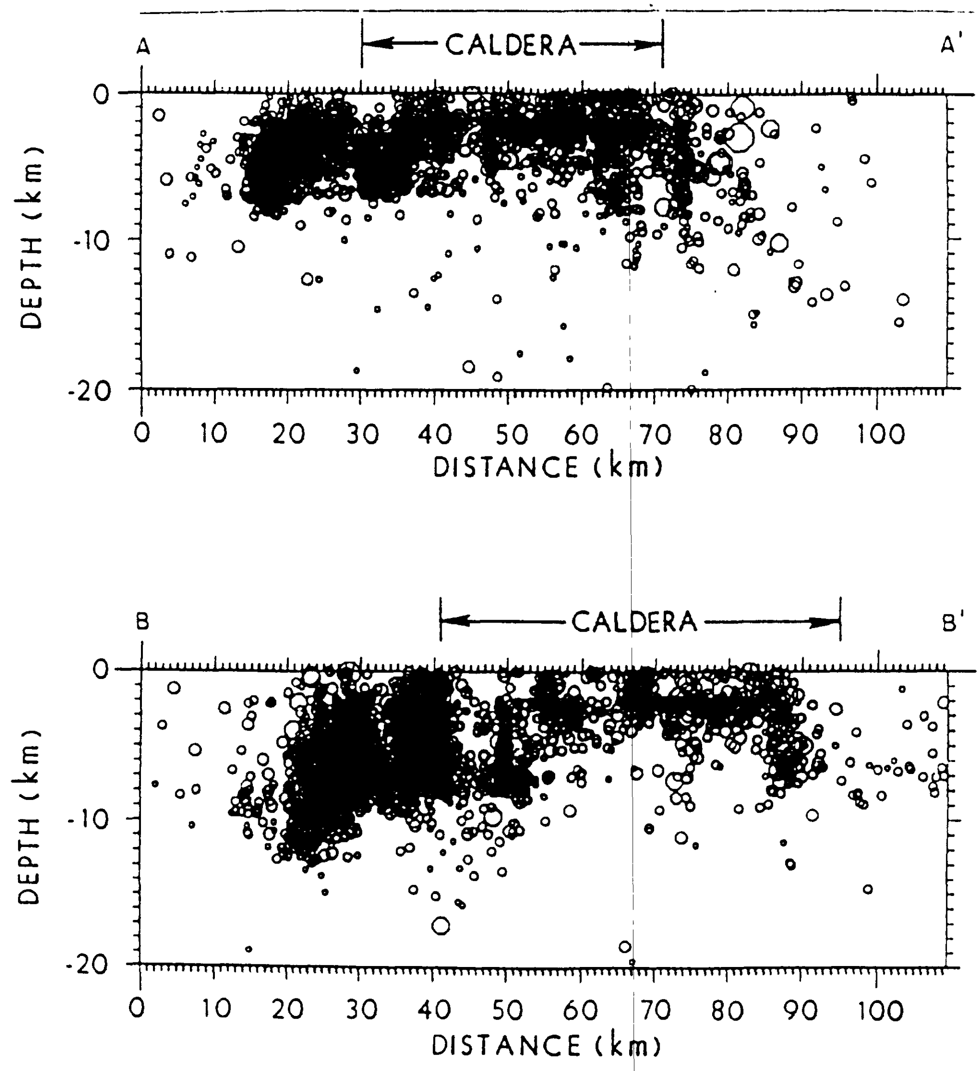

Figure 17. Cross sections of seismicity beneath the Yellowstone region, keyed to Figure 16. Data correspond to that described for Figure 16, but extended to all shocks of $M_L > 0$. Plots include only hypocenters within 10 km of the planes of section for which there were more than 7 recording stations, a map distance to the nearest recording station less than twice the focal depth, and an azimuthal gap in station coverage less than 180°.

east-west band of seismicity that passes through the Hebgen Lake region. The 1959 Hebgen lake main shock (discussed earlier) occurred within about 30 km of the Yellowstone caldera. The earthquake may have resulted from unusual lithospheric uplift and viscoelastic relaxation associated with the Yellowstone hot spot (Reilinger, 1986).

Along the northwest side of the SRP, Figure 15 shows a pronounced northwest alignment of epicenters between Mackay and Challis, which is aftershock activity of the 1983 Borah Peak earthquake on the Lost River fault. This pattern contrasts with the scatter of what we have called background seismicity elsewhere in the central ISB. The "turning on" of earthquakes on the Lost River fault emphasizes the relative seismic quiescence of the neighboring Lemhi and Beaverhead faults to the northeast. All three faults are part of a domain of active, latest Quaternary basin-range normal faulting northwest of the SRP (Scott and others, 1985). Hence, the paucity of earthquakes between the Lost River fault and the Idaho-Montana border marks another important seismic gap in the central ISB. Seismic surveillance of this region by the Idaho National Engineering Laboratory (King and others, 1987) shows small-magnitude earthquake activity near the central part of the Beaverhead fault, but very minimal microseismicity along the Lemhi fault. Central Idaho west of the Borah Peak earthquake zone (Fig. 15) is characterized by diffuse earthquake activity (Dewey, 1987; Smith and Sbar, 1974), earthquake swarms (Pennington and others, 1974; Smith, 1977), and extensive hot spring activity. Microearthquake studies reported by Smith (1977) suggest maximum focal depths of 10 to 15 km.

Northern ISB

The northern ISB, as we have delimited it north of the Hebgen Lake–Yellowstone region, lies entirely within western Montana (Fig. 1). The region has had no historical surface faulting, and the largest historical earthquake reached magnitude 6¼ (Fig. 9, Table 2). Here, we expand upon a recent summary of seismicity and faulting in western Montana–eastern Idaho by Stickney and Bartholomew (1987) to describe some detailed aspects of the northern ISB. For illustration, Figure 18 combines available information on Cenozoic faulting with a representative seven-year sample of instrumental seismicity from the Montana regional seismic network (M, Fig. 8B).

The area of Figure 18 contains at least three fundamentally different domains of Cenozoic basin-bounding extensional faults (see Fig. 3 for names): (1) large north-northwest–trending faults in northwestern Montana, (2) west-northwest–trending faults marking the Lewis and Clark Zone (LCZ), and (3) faults of variable trend south of the LCZ. Significantly, none of these domains include faulting of Holocene age; Holocene faulting in Montana is restricted to a belt along the northwest flank of the SRP (the Centennial Tectonic Belt of Stickney and Bartholomew, 1987), which we have described as part of the central ISB. Large faults in northwestern Montana have exerted strong control on Cenozoic topography, probably including significant Quaternary displacements (Pardee, 1950), but the current seismic potential of these faults is uncertain (Qamar and others, 1982). The LCZ is a pre-Cenozoic structural lineament about 400 km long, perhaps dating from the Proterozoic, that has been interpreted to reflect a fundamental intraplate boundary (see Stickney and Bartholomew, 1987). Major transcurrent shearing has been postulated for the LCZ, but the zone is not noted for any such slip in late Cenozoic time (Eardley, 1962). In its modern expression, the LCZ is considered to be a transitional zone, up to 50 km wide, that "divides a region of uniformly northwest-trending Laramide thrusts and folds and Tertiary normal faults to the north from a

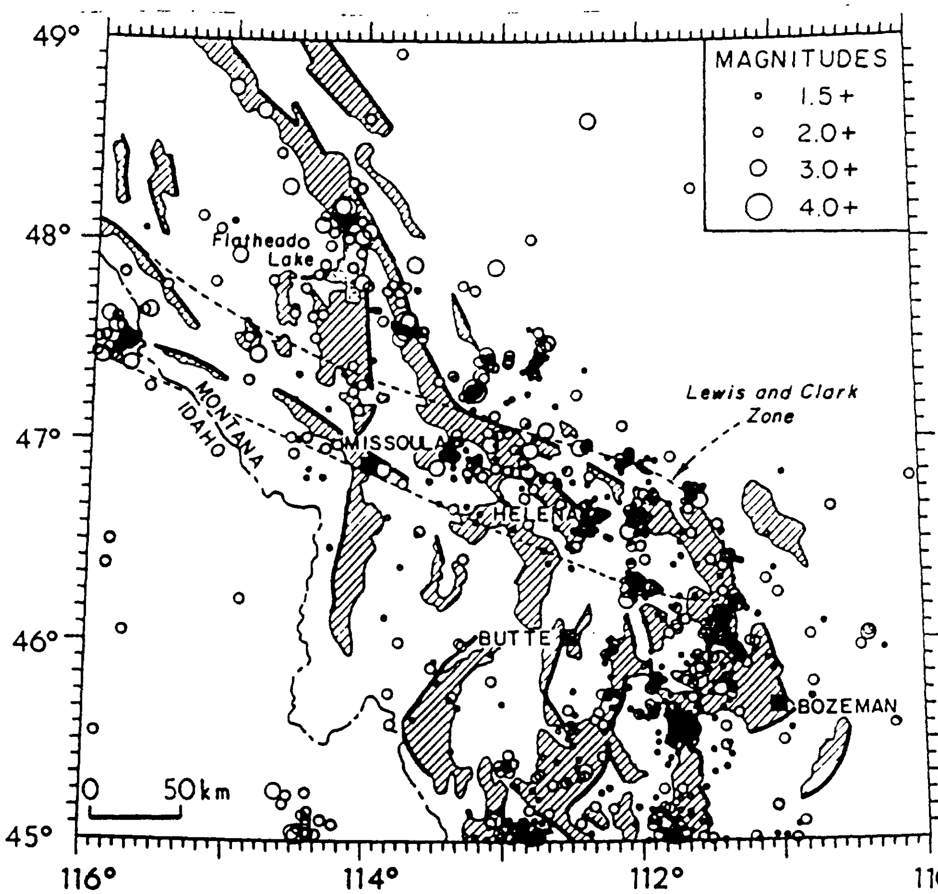

Figure 18. Seismicity map of the northern Intermountain seismic belt. Map shows all earthquakes of magnitude 1.5 and greater located by the Montana Bureau of Mines and Geology from 1982 through 1987 (M. C. Stickney, personal communication, 1990). Base map shows Cenozoic basins (hachured), basin-bounding extensional faults (heavy lines), and the outline of the Lewis and Clark Zone (short-dashed line) after Stickney and Bartholomew (1987).

region of batholithic intrusions and shorter Basin and Range structures with diverse trends to the south" (Stickney and Bartholomew, 1987).

The background seismicity shown in Figure 18 defines a belt about 200 km wide that follows the regional northwest trend of extensional basins in western Montana without definite association with the mapped faults. We describe features of this seismicity in each of the three domains of faulting outlined above. In northwestern Montana, historical seismicity has not exceeded magnitude 5.5 (Fig. 9) and has been dominant in the vicinity of Flathead Lake (Figs. 1, 9, 18), where significant earthquake swarms occurred in 1945, 1952, 1964, 1969, 1971, and 1975 (Qamar and others, 1982). A possible relation between seismicity and reservoir loading at Flathead Lake has been suggested by Dunphy (1972). Results from short-term seismographic recording near Flathead Lake in 1971 by Stevenson (1976; see also Sbar and others, 1972) showed clustering of small earthquakes on the west side of the lake, with focal depths extending from 5 to 12 km depth and foci defining a planar zone dipping 70° to the east-northeast. Diverse focal mechanisms in the vicinity of Flathead Lake appear mostly to reflect east-west to northwest-southeast extension (Qamar and others, 1982). Seismicity near Flathead Lake roughly marks the northernmost extent of the ISB as it dies out toward the Canadian border (Figs. 1 and 18). Coincidentally, Flathead Lake lies at the southern end of the Rocky Mountain trench, a large pre–basin-range graben that extends more than 800 km northwestward into Canada (Eardley, 1962; Pardee, 1950).

Stickney and Bartholomew (1987) attach great importance to the LCZ as the northern boundary of the active extensional regime of the Montana-Idaho part of the Basin and Range province. All identified late Quaternary faults and all historical earthquakes greater than magnitude 5.5 in the northern ISB are south of, or are included within, the LCZ. Seismicity shown in Figure 18 within the LCZ includes relatively dense clusters of earthquakes from east of Missoula to the southeastern end of the LCZ and a distinct cluster of events at the northwest end of the LCZ in the figure. The latter include both rockbursts related to deep mining near Wallace, Idaho, and local tectonic earthquakes (Stickney and Bartholomew, 1987). We earlier discussed the destructive swarm earthquakes of 1935–1936 near Helena. Earthquake studies reported by Friedline and others (1976) near Helena indicate earthquake clustering along a N50°W trend with focal-depth maxima of 15 to 20 km. Focal mechanisms for this area imply normal to lateral slip with consistent northeast-trending T-axes (Friedline and others, 1976; Stickney and Bartholomew, 1987; Doser, 1989a).

Background seismicity south of the LCZ in Figure 18 is concentrated in a northerly-trending zone about 100 km wide between the eastern part of the LCZ and the Hebgen Lake–Yellowstone region to the south. Recall that the Helena-Bozeman area has had the largest historical earthquakes in the northern ISB (Fig. 9). Densely clustered seismicity about 50 km northwest of Bozeman is in the Clarkston Valley area, one of the most persistently active earthquake zones in the northern ISB since the occurrence of the M = 6¾ Clarkston Valley earthquake in 1925 (Stickney and Bartholomew, 1987). Special studies of seismic activity in the Clarkston Valley area have been reported by Qamar and Hawley (1979), who document relatively shallow (<10 km) focal depths and both normal and strike-slip focal mechanisms with T-axes trending 55° to 95°—different from the trend of 127° for the 1925 main-shock mechanism (Doser, 1989a).

INDUCED SEISMICITY

Human activities have demonstrably triggered small to moderate earthquakes ($M \leqslant 5.0$) in the Intermountain region, both within the main active zone of the ISB and in marginal, less seismically active areas. This includes documented cases of each of the three principal types of induced seismicity—associated, respectively, with reservoir impoundment, fluid injection, and mining. We proceed to summarize examples (see also Dewey and others, 1989, Table 1), referring to localities identified in Figure 3.

Reservoir-induced seismicity

The pre-eminent case of reservoir-induced seismicity (RIS) in the Intermountain region is that associated with Lake Mead along the Arizona-Nevada border—impounded by the 221-m-high Boulder Dam (now Hoover Dam) on the Colorado River in the mid-1930s. Locally felt earthquakes began soon after filling of what was then the world's largest reservoir, and a magnitude 5.0 shock occurred in 1939 about a year after the reservoir reached 80 percent capacity (Carder, 1970). Observations by Carder (1945) represent the first instance in which the phenomenon of RIS was recognized. By the mid-1970s more than 10,000 local earthquakes had been recorded near Lake Mead, with an irregular epicentral distribution inferred to be influenced chiefly by the variable permeability of sub-reservoir sedimentary rocks rather than by lateral differences in mass loading (Anderson and Laney, 1975). Rogers and Lee (1976; see also Rogers and others, this volume) describe results of detailed seismographic monitoring at Lake Mead during 1972–1973 and suggest observed RIS there is caused by small increases in pore pressure along existing faults.

Reviews of RIS on a worldwide basis (e.g., Simpson, 1976; Gupta and Rastogi, 1976; Gupta, 1985) typically mention five other likely or possible cases of RIS in the Intermountain region. These include: (1) Kerr Dam (Flathead Lake), Montana, one of at least 14 worldwide dams associated with earthquakes in the magnitude 4 range (Gupta, 1985; note that earthquakes of magnitude 4.6 and 4.9 occurred near Kerr Dam 6 and 13 years, respectively, after impoundment in 1958); (2) Glen Canyon (Lake Powell) and (3) Flaming Gorge dams in Utah, where there is uncertain evidence for impoundment-related *decreases* in seismicity (Simpson, 1976); and possible cases of RIS at (4) Clark Canyon Dam, Montana, and Palisades Dam, Idaho (Gupta, 1985; Simpson, 1976). The common occurrence of earthquake swarms throughout the region immediately surrounding Palisades Reservoir confounds arguments for causally relating such seismicity with the reservoir (see review by Piety and others, 1986). There, as elsewhere in the Intermountain region, adequacy of seismographic control is a critical issue in correlating small-magnitude earthquakes with reservoir impoundment. LaForge (1988), for example, rigorously tested for RIS in connection with 14 dams, 38 to 81 m high, built by the U.S. Bureau of Reclamation in north-central Utah and found no evidence for RIS. However, all but one of the reservoirs was filled before 1967 when seismographic control in the region was still relatively poor.

Fluid injection

After the U.S. Army accidentally triggered earthquakes by deep fluid injection near Denver, Colorado, in the early 1960s, the U.S. Geological Survey carried out a controlled experiment in the Rangely oil field in northwestern Colorado to study the effects of pore-pressure changes at depth on the triggering of small earthquakes (Raleigh and others, 1972). Water injection into a sandstone reservoir about 2 km deep began in late 1957 at Rangely for secondary oil recovery. The spatial coincidence of earthquakes ($M \leqslant 4.5$) with the Rangely field can be shown at least as early as November 1962 when the UBO array (Fig. 8A), 50 to 80 km to the west-northwest, began operating (Munson, 1970). The detailed experiment carried out by the U.S. Geological Survey at Rangely in 1969–1970 successfully measured in situ stress state and showed how lowering and raising fluid pressures in a seismically active zone at depth could control the occurrence of earthquakes (Raleigh and others, 1972).

Arabasz (1984; see also Arabasz and Julander, 1986, Fig. 8) describes another possible case of seismicity related to fluid injection in the Intermountain region. In mid-1982 an "acid-breakdown hydrofrac" was made in a wellbore at a depth of about 5 km (Chevron U.S.A. #1 Chriss Canyon, total depth = 5,344 m) in the vicinity of the southern Wasatch fault. Hypocentral clustering of two earthquake swarms ($M_L \leqslant 2.1$) two to three months later, within a few kilometers of the wellbore and with distance-delay times consistent with fluid diffusion, suggests triggering by the fluid injection. Geothermal steam production has been monitored for induced seismicity at two sites in the Intermountain region (Zandt and others, 1982): Roosevelt Hot Springs–Cove Fort, Utah (ca. 38°30′N, 112°45′W), and Raft River, Idaho (42°06′N, 113°23′W), but to our knowledge no significant induced seismicity has occurred.

Mining-related seismicity

Mining-related seismicity has been specially investigated in two parts of the Intermountain region—the northern and northwestern Colorado Plateau, chiefly in association with underground coal mining, and near Wallace, Idaho, in association with deep vein mines of the Couer d'Alene mining district. In the latter area, seismographic studies have focused on rockbursts (e.g., McLaughlin and others, 1976); however, sizable shocks up to magnitude 4 suggest tectonic stress release as well (Stickney and Bartholomew, 1987).

Seismicity in east-central Utah defines an inverted U-shaped pattern (Figs. 1 and 12B) coinciding with areas of extensive underground coal mining along an arcuate erosional escarpment of the eastern Wasatch Plateau and Book Cliffs. The association of both rockbursts and earthquakes ($M_L \leqslant 4.5$) with sites of major coal extraction in this area has been evident since the late 1950s. Wong and Humphrey (1989) and Williams and Arabasz (1989; see also Smith and others, 1974) give good overviews of varied seismological investigations indicating that: (1) much of the seismicity appears to be mining induced, resulting from stress redistribution from both room-and-pillar and longwall coal extraction in mine workings down to 900 m below the surface; (2) abundant seismicity occurs *beneath* the mines to depths of 2 to 3 km; (3) time-varying rates of extraction at individual mines have influenced seismicity changes detected by regional seismic monitoring; and (4) source mechanisms appear variously to reflect extensional subsidence above mine workings and a mixture, at and below mine level, of seismic slip on prestressed reverse faults and possibly non-double-couple, implosional failures.

Figure 19A shows abundant submine events located in the Gentry Mountain area of the eastern Wasatch Plateau. Nearly all these events were recorded with ubiquitous dilatational first motions, inferred to be caused by implosional failure (Wong and others, 1989). Companion results from the nearby East Mountain area (Williams and Arabasz, 1989; Fig. 19B) also show reliably-located submine events, but mostly with reverse-faulting mechanisms (normal-faulting solutions 1 and 2 are for earthquakes west of the mining area beneath an active graben). About a third of the seismic events recorded in the East Mountain area were of the enigmatic type with all dilatational first motions. Given inadequate focal-depth resolution for many of these shallow events, Williams and Arabasz (1989) found that if these events were constrained to occur at mine level, their first-motion distributions were indeed incompatible with a double-couple source mechanism. However, the same first-motion observations could be fit with double-couple normal-faulting solutions if the sources were *above* mine level, perhaps reflecting overburden subsidence.

We refer the reader to Wong and Humphrey (1989) for a review of other varied work on mining-induced seismicity in the Colorado Plateau. This includes (1) studies by the U.S. Geological Survey of seismicity induced by underground coal mining near Somerset in western Colorado and (2) studies by Wong and coworkers of seismicity induced by potash mining in the north-central Colorado Plateau involving brine extraction from a previous room-and-pillar mine at 1 km depth.

PATTERNS OF EARTHQUAKE OCCURRENCE

Observations about patterns of earthquake occurrence in the ISB are fundamental both for scientific understanding of earthquake behavior in the region as well as for basic evaluations of

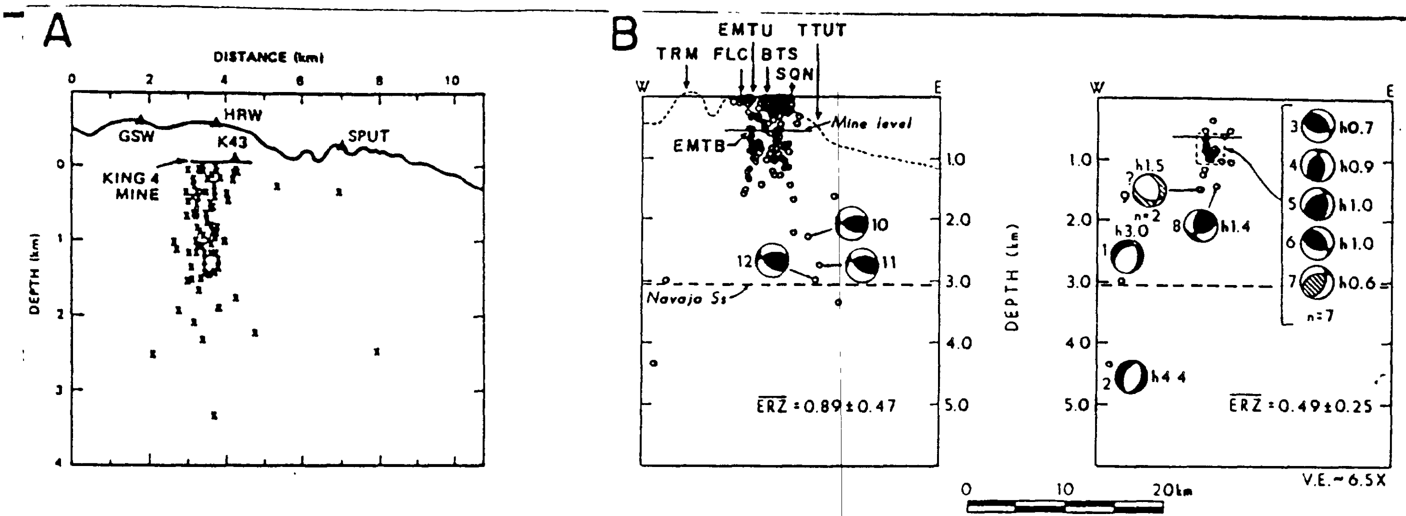

Figure 19. Mining-related seismicity in the eastern Wasatch Plateau, Utah. A, Cross section from a special study in the Gentry Mountain area (from Wong and others, 1989) showing ground surface (heavy line), seismic events (x's) within 500 m of the plane of section, and closest seismographs (triangles). B, Cross sections from the East Mountain area, about 20 km south of Gentry Mountain (from Williams and Arabasz, 1989) showing ground surface (short-dashed line), seismic events (small circles) within 5 km of the plane of section, seismograph locations (letter codes) within 1 km of the plane of section, and schematic equatorial-plane projection of 12 focal mechanisms—dilatational quadrants are white and compressional quadrants either black (for single-event solutions) or hachured (for composite solutions); the left panel of B includes events located with standard errors in focal depth (ERZ) less than 2.0 km; the right panel includes only the "best" located events meeting rigorous criteria for reliability of focal depth, h.

earthquake hazards and risk. In this section we briefly summarize relevant information on (1) earthquake swarms, (2) earthquake recurrence, and (3) the space-time distribution of earthquakes in the ISB.

Earthquake swarms

Earthquake swarm activity, the clustering of earthquakes of similar size in space and time without an outstanding main shock, is a common feature in parts of the ISB. Such earthquake swarms tend to occur in and near areas of Quaternary volcanism or high heat flow. Smith and Sbar (1974) present a good general discussion, describing notable areas of historical swarm activity in the ISB. These include: Flathead Lake and Helena, Montana, in the northern ISB; Yellowstone, central Idaho, and southeastern Idaho in the central ISB; and areas in western and southwestern Utah in the southern ISB. We have already described sources of information for most of these areas in preceding sections. Additional information on earthquake swarm activity is given by Arabasz and Julander (1986) for southwestern Utah and by Piety and others (1986) for southeastern Idaho. The destructive earthquake swarm near Helena, Montana, in 1935–1936 (described earlier) that included shocks of magnitude 6 and 6¼ (Table 2) is the most outstanding example of swarm seismicity in the ISB. Elsewhere in the ISB, the largest earthquakes in individual swarms have been in the upper magnitude 4 range or smaller.

Earthquake recurrence in the ISB

Earthquake recurrence specifies the distribution of earthquake sizes (e.g., magnitudes or intensities) and their frequency of occurrence on a fault or in a specified area. Here, we basically want to describe how often moderate to large earthquakes occur in the ISB and point the interested reader to more specific information for recurrence modeling. Modern methods for quantifying earthquake recurrence using information from observational seismology and from geologic studies of the age, frequency, and rupture characteristics of prehistoric earthquakes are summarized by McGuire and Arabasz (1990) and Schwartz and Coppersmith (1986) (see also Doser and Smith, 1982).

The most commonly used relations to describe the relative number of earthquakes as a function of size is the well-known Gutenberg-Richter relation (Richter, 1958): $\log_{10} N(m) = a - bm$, where $N(m)$ is the number of earthquakes of magnitude m or greater per unit time (and ideally per unit area) and a and b are constants. The average inter-event time or recurrence interval for earthquakes of a particular magnitude m or greater is given by $1/N(m)$. There is growing recognition that the Gutenberg-Richter relation may appropriately apply to a region but not necessarily to an individual fault (e.g., Schwartz and Coppersmith, 1986). It is also recognized that the historical and instrumental earthquake record cannot confidently be extrapolated to estimate the frequency of occurrence of large surface-faulting earthquakes in the

ISB and that information from late Quaternary faulting is therefore essential (Schwartz and Coppersmith, 1984; Arabasz and others, 1987). Thus, historical and instrumental records of seismicity in the ISB are important for modeling the recurrence of earthquake sizes up to the threshold of surface faulting, and paleoseismology is needed for estimating how often large surface-faulting earthquakes will occur.

A rough estimate of the recurrence interval of sizable earthquakes for the *whole* ISB can be made from Table 2. Neglecting aftershocks and secondary events, 27 independent main shocks of approximate magnitude 5.5 and greater occurred in the ISB from 1900 through 1985, which gives an average inter-event time of 3 years for such shocks somewhere in the ISB. Corresponding estimates from Table 2 for thresholds of approximate magnitude 6.0 and 6.5 are about 6 years and 17 years, respectively. For earthquakes of magnitude 7.0 or greater, we simply note that two such events occurred during the 85-year observation period, insufficient for meaningful modeling of inter-event times. Recurrence intervals vary, of course, with the particular region and the total area being considered. For example, for the 85,000 km^2 of the Wasatch Front area (Fig. 12), rigorous recurrence modeling of instrumental seismicity for 1962 through 1985 (Arabasz and others, 1987) yields average recurrence intervals of 24 years for $M_L \geqslant 5.5$, 54 years for $M_L \geqslant 6.0$, 120 years (extrapolated) for $M_L \geqslant 6.5$, and 280 years (extrapolated) for $M_L \geqslant 7.0$. For other examples of recurrence modeling in parts of the ISB, we refer the reader to Youngs and others (1987), Doser and Smith (1982), Piety and others (1986), Stickney and Bartholomew (1987), and Algermissen and others (1982). Comparison of seismicity parameters among these and other published reports for the ISB must be made with care, checking whether the parameters were determined with sufficient rigor, whether the parameters are intrinsically comparable, and whether the parameters describe the occurrence only of independent main shocks or of all earthquakes (see McGuire and Arabasz, 1990).

Average slip rates on normal faults in the ISB are one to two orders of magnitude lower than for those on major plate-boundary faults, typically being about 1 mm/yr for the most active faults like the Wasatch fault (Schwartz and Coppersmith, 1984; Machette and others, 1987) and the Teton fault (Byrd and Smith, 1990), and a few tenths of a millimeter or less on faults elsewhere in the region (e.g., Schwartz, 1987; Youngs and others, 1987; Scott and others, 1985; Stickney and Bartholomew, 1987). Corresponding recurrence intervals for surface rupture on an individual fault segment are about 2,000 yr on the most active parts of the Wasatch fault, are uncertain on the Teton fault, and are typically several thousand or tens of thousands of years on other faults. The issue of uniform versus time-varying recurrence (see Schwartz, 1988) has become an important consideration for estimating expected rates of occurrence of large earthquakes in the ISB. On the Wasatch fault, for example, the record of surface-faulting earthquakes during the past 6,000 yr leads to an average recurrence interval of 415 yr for a large surface-faulting earthquake somewhere on the fault, but an accelerated rate of faulting between about 400 and 1,500 yr ago implies such an earthquake once every 220 yr (Machette and others, 1989). Comparably detailed paleoseismological data do not exist for other faults in the ISB.

How do rates of intraplate faulting and earthquake activity in the ISB compare to those along the North American plate boundary in California? Recurrence intervals of thousands of years for surface rupture on individual fault segments in the ISB compare to much shorter intervals of hundreds of years for large earthquakes on the most active parts of the San Andreas fault system (Schwartz and Coppersmith, 1984). Maximum-size earthquakes of about magnitude 7¾ in the ISB compare to a value of about 8½ on the San Andreas fault. We remarked earlier that deformation rates for most of the intraplate ISB are one to two orders of magnitude lower than along the western North America plate boundary. Comparison between seismicity in the ISB and California can be made using seismicity rates determined in a uniform way by Algermissen and others (1982, Table 1) for source zones throughout the United States, normalizing those values per unit area. The mean rate of earthquakes equivalent in size to MMI = V (about magnitude 4) per year per 1,000 km^2 is approximately 8×10^{-2} for the 41 seismic source zones depicted by Algermissen and others (1982, Fig. 2) in the main seismically active part of California. The mean value of that same rate for their seismic zones making up the main ISB is approximately 2×10^{-2}. Thus, normalized seismicity in the ISB, on average, is lower by about a factor of 4 compared to that along the plate boundary in California.

Space-time patterns of seismicity in the ISB

To get an overview of variations of earthquake activity along the ISB as a function of space and time, as captured by the DNAG catalog, we have plotted those data in a conventional space-time format and show the results in Figure 20. The same DNAG catalog data used for Figure 1 were sorted for the four sample areas shown in Figure 20A, prescribing a magnitude threshold of 3.0 and the time period from 1930 through 1985. We chose 1930, judging that epicentral precision had become sufficient by that time to make meaningful spatial comparisons, and the date precedes the occurrence of sizable earthquakes in the ISB in the mid-1930s and 1940s. The sorted earthquakes are plotted in Figure 20B with latitude as the space coordinate, given the general north-south trend of the ISB, recognizing the limitation that the northern and southernmost parts of the ISB trend obliquely to the latitude ordinate. Data for central Idaho had to be excluded to prevent confusion in projection.

As usual, there are evident artifacts that must be accounted for in space-time plots of this type, the most obvious being those due to catalog incompleteness. Increases in numbers of earthquakes in the early 1960s and locally in the mid-1970s correspond to improvements in seismographic coverage that we described in the section on instrumental recording and seismic networks. Despite recognizable problems, Figure 20 reveals some

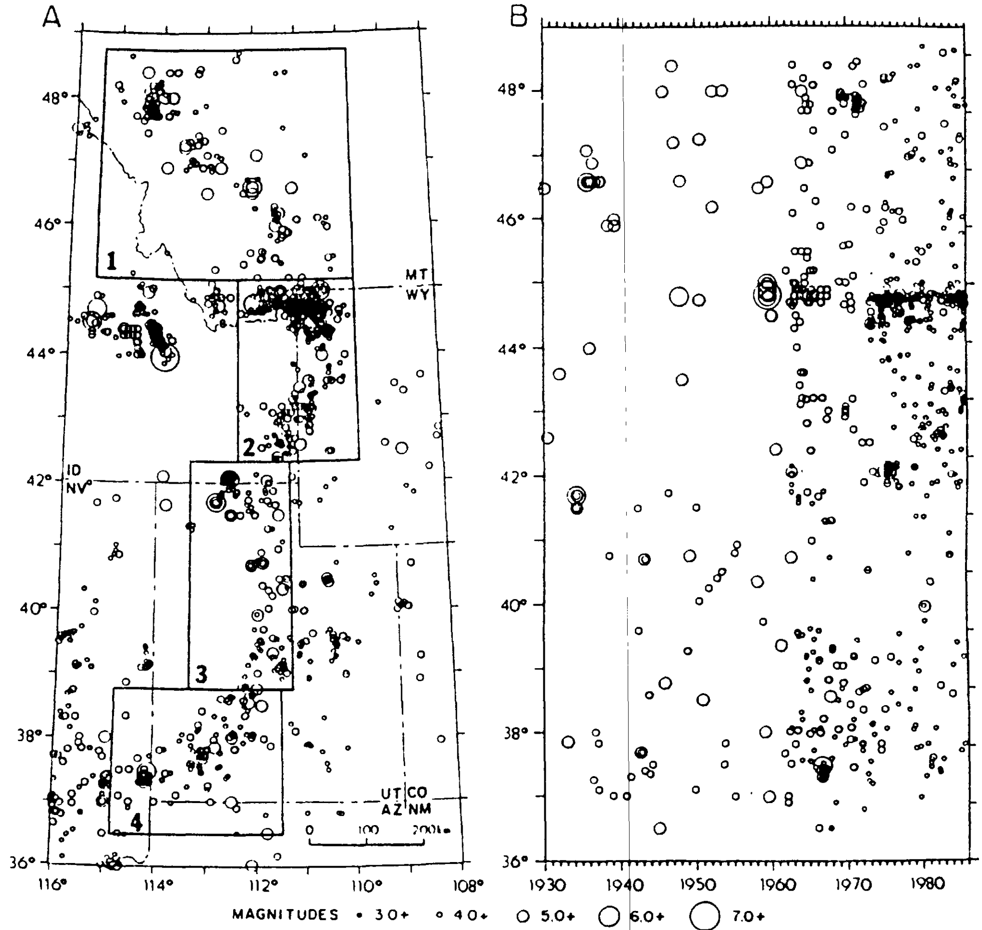

Figure 20. Space-time seismicity of the Intermountain seismic belt, 1930 to 1985. A, Map showing earthquakes from Figure 1 of magnitude 3.0 and greater since 1930, together with sample boxes used for the space-time plot in B. B, Space-time plot of earthquakes as a function of time (abscissa) and latitude (ordinate), keyed to A.

important observations that warrant attention. We proceed from north to south, using the boxes numbered from 1 to 4 for reference.

For the northern ISB, the space-time projection of earthquakes in Box 1 shows an understandable increase in smaller-magnitude earthquakes after about the mid-1960s, but there is a noticeable decrease in the number of larger earthquakes ($M \geqslant 5$) after that time, compared to earlier decades. Seismicity in the Hebgen Lake–Yellowstone region dominates the picture for the part of the central ISB bounded by Box 2. The projection of the Hebgen Lake–Yellowstone seismicity reveals the most intense earthquake activity in the ISB. There was a marked increase in the detection of small earthquakes following the installation of the Yellowstone seismic network in 1973, but similarly sensitive networks elsewhere in the ISB do not show the same level of seismicity. The $M_S = 7.5$ Hebgen Lake earthquake of 1959 is the largest shock at that latitude. Nearby prior occurrence of the $M = 6\frac{1}{4}$ Virginia City earthquake 12 years earlier in 1947 suggests that earthquake may have been a preshock (Doser, 1989a). Both the map and space-time plot emphasize the existence of the Teton seismic gap, more than 50 km long, immediately south of the Yellowstone region. As earlier described, there is a similarly prominent gap northwest of the SRP between the Borah Peak earthquake zone and the Idaho-Montana border (Fig. 20A).

Box 3 essentially covers the Wasatch Front area. The $M_L = 6.0$ Pocatello Valley earthquake of 1975, at about 42°N, was preceded by seismic quiescence within 50 km that began 6.4 years before the main shock (Arabasz and Smith, 1981). One of the most striking features of the space-time projection of the earthquakes of Box 3 is the sparseness of earthquakes ($M \geqslant 3.0$) since the mid-1960s along a zone more than 200 km long between about 39.5°N and 41.5°N, corresponding to the most geologically active part of the Wasatch fault (Fig. 12). Note that there has been effective seismographic coverage of the Wasatch Front area since the early 1960s (Fig. 8B). Smith (1972) first noted the relative contemporary quiescence of the Wasatch Front area compared to neighboring segments of the ISB, and Arabasz and Smith (1981, Fig. 8; see also Griscom, 1980) portrayed the relative quiescence in space-time view. The apparent decrease in background seismicity in the Wasatch Front area beginning in the mid-1960s compared to prior decades was statistically tested by Arabasz (1984), who found that one could not reject at the 95-percent confidence level the hypothesis that the apparent decrease was simply random. The pattern of earthquake clustering in the Wasatch Front area, however, does have significant features. Based on a statistical analysis of the Utah earthquake catalog for 1962 through 1985 by Shimizu (1987), Veneziano and others (1987) observed that, compared to earthquake behavior in neighboring areas, there is a distinct paucity of secondary events, relative to the number of main events, in the vicinity of the Wasatch fault between 39.5°N and 41.5°N. We refer the reader to Arabasz and others (1987, Fig. 17) for a detailed space-time plot of microseismicity within a 30-km-wide zone along the Wasatch fault for 1962 through 1986.

The space-time plot of earthquakes in Box 4 at the southern end of the ISB shows a somewhat similar pattern to that for Box 1. Despite a marked improvement in seismographic coverage for the area of Box 4 in the mid-1970s, background seismicity has been noticeably lower since that time, compared to activity during the 1962 through 1975 period. Earthquake sizes in the Utah region have been measured instrumentally in a uniform way since 1962 (Arabasz and others, 1979). Although not rigorously analyzed here, the space-time overview of the entire ISB in Figure 20 points out general patterns that need to be addressed. The occurrence of the 1983 Borah Peak earthquake in an area of prior low seismicity and prehistoric surface faulting, which should have been recognized as a seismic gap, serves as a useful reminder.

CONCLUDING REMARKS

The mechanics and subsurface geometry of normal faulting are topics of great current interest and importance, especially for assessing ground deformation and peak ground motions associated with large normal-faulting earthquakes. The observation in parts of the Basin and Range province of steep planar faults (dips > 45°) associated with large normal-faulting earthquakes in the same structural setting where late Tertiary to Quaternary low-angle and listric normal faults are present in the subsurface poses a quandary. In this final section, we discuss observations and hypotheses relating to large normal-faulting earthquakes and to the diffuse background seismicity that dominates the earthquake record of the Intermountain region. We also consider implications of earthquake focal depths for rheology and the possible influence of the Yellowstone hot spot on the seismotectonics of the ISB, including its influence on Quaternary faulting of the SRP and surrounding region. Discussion of earthquakes in the Yellowstone region also provides insights into the relation between magma transport and seismicity.

Correlation of seismicity and geologic structure

A recurring observation in the ISB is the lack of distinct correlation between scattered background seismicity and mapped Cenozoic faulting. Numerous descriptions of background seismicity in the region published during the last decade include remarks to this effect. The problematic correlation has been discussed at length for the southern ISB by Arabasz and Julander (1986; see also Arabasz and Smith, 1981; Zoback, 1983; Smith and Bruhn, 1984). As outlined by Arabasz and Julander (1986), and generalized to the ISB as a whole, the basic problems include: (1) uncertain subsurface structure, which typically is more complex along the main seismic belt than is apparent from the surface geology, commonly because of the superposition of basin-range faulting upon older thrust-belt structure; (2) observations of discordance between surface fault patterns and seismic slip at depth; (3) limited opportunity to observe large-scale seismic slip because of only three cases of historic surface faulting; and (4) inadequate hypocentral resolution commonly resulting from regional seismic monitoring. In order to correlate seismicity with structure, there is a critical need (aptly illustrated by Fig. 14) for local seismographic control, especially for good focal-depth resolution, sufficient seismicity for defining the spatial geometry of active structures, and reliable focal mechanisms for correlating observed seismicity with fault geometry and the sense of slip. Focal mechanisms also allow assessment of the principal stress directions.

On the basis of special earthquake studies in mostly the southern ISB, a working hypothesis was offered by Arabasz (1984; see also Arabasz and Julander, 1986) to explain observations of diffuse background seismicity. Background seismicity, it was suggested, is fundamentally influenced by variable mechanical behavior and internal structure of individual plates within the seismogenic upper crust. Diffuse epicentral patterns appear to result from the superposition of relatively intense shallow seismicity within upper-crustal plates and less frequent background earthquakes at greater depth. Favorable conditions for block-interior rather than block-boundary microseismic slip may also contribute to the epicentral scatter. Some aspects of the working hypothesis are shown schematically in Figure 21A, which depicts (following Arabasz and Julander, 1986): (a) a predominance locally of seismicity within a lower plate; (b) nucleation of a large normal-faulting earthquake near the base of the seismogenic layer, hypothetically on an old thrust ramp, and with linkage to a shallow structure; (d) occurrence of a moderate-sized earthquake and aftershocks on a secondary fault where an underlying detachment restricts deformation to the upper plate; (e) diffuse block-interior microseismicity predominating within an upper plate—perhaps responding to extension enhanced by gravitational backsliding on an underlying detachment; and (f) diffuse block-interior microseismicity within a lower plate where frequency of occurrence is markedly lower than in the overlying plate.

While Figure 21A suitably illustrates many features of background seismicity in the southern ISB, particularly central Utah, it is not adequately general for the whole ISB. One shortcoming of the sketch in Figure 21A is that it does not explicitly depict the spatial relation of seismicity to Precambrian basement. In some parts of the ISB, Precambrian basement was faulted by

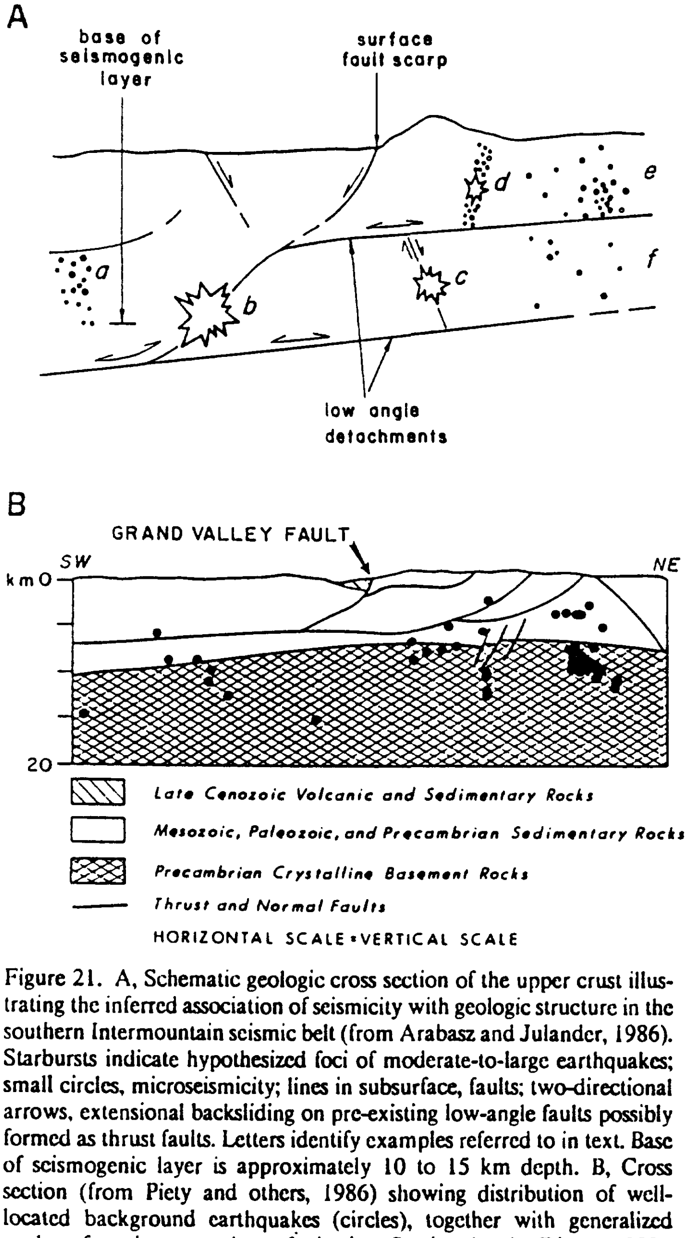

Figure 21. A, Schematic geologic cross section of the upper crust illustrating the inferred association of seismicity with geologic structure in the southern Intermountain seismic belt (from Arabasz and Julander, 1986). Starbursts indicate hypothesized foci of moderate-to-large earthquakes; small circles, microseismicity; lines in subsurface, faults; two-directional arrows, extensional backsliding on pre-existing low-angle faults possibly formed as thrust faults. Letters identify examples referred to in text. Base of seismogenic layer is approximately 10 to 15 km depth. B, Cross section (from Piety and others, 1986) showing distribution of well-located background earthquakes (circles), together with generalized geology from interpretations of seismic-reflection data by Dixon (1982), across the Grand Valley fault in the central Intermountain seismic belt. Line of section trends N60°E and crosses the northern end of Palisades Reservoir (Fig. 3).

pre-Neogene thrusting and is allochthonous, as in parts of the Wasatch Front area (Smith and Bruhn, 1984); in others areas, such as that studied by Piety and others (1986) in the central ISB, Precambrian crystalline basement rocks lie beneath a basal detachment and are autochthonous. For the latter area we noted that relatively abundant background microseismicity was located by Piety and others (1986), as shown in Figure 21B, within the Precambrian crystalline basement, and they observed no marked change in the vertical distribution of foci that could be associated with the basal detachment.

The present thickness of sedimentary cover rocks, the extent of involvement of Precambrian basement in older compressional deformation and subsequent extension, and whether normal faulting penetrates the entire crust (Zandt and Owens, 1980) all are likely to be important local factors governing the pattern of background seismicity. The influence of these respective factors throughout the ISB is not yet understood, but there are accumulating observations, such as seismic slip on discrete Precambrian basement faults at depth (Fig. 14), the broad involvement of Precambrian basement in contemporary extension (Fig. 21B), and perhaps the regional, flexural(?) deformation of Precambrian basement on the eastern, footwall side of the Wasatch fault (Fig. 12B).

Earthquake focal depths and rheology

Since the mid-1970s, accurate hypocenter data have been acquired by regional and portable seismic networks in the Intermountain region that permit the construction of reliable focal-depth histograms. Using some of these data, Sibson (1982) and Smith and Bruhn (1984) hypothesized seismogenic models based on theoretical depths for peaks in maximum shear stress at the boundary between the brittle upper crust and a quasi-plastic layer. These models in a general way account for the maximum depths of nucleation of large normal faulting earthquakes and for the maximum depths of background seismicity, corresponding to the base of the seismogenic layer. The models involve a temperature-dependent, depth-varying power law for creep combined with a linear brittle-behavior criterion (see representative plot in Fig. 22). In Smith and Bruhn's (1984) model for extensional normal-faulting regimes, the maximum focal depths of large normal-faulting earthquakes correlate approximately with the 80th percentile of focal depths for smaller background earthquakes, similar to the findings of Sibson (1982) for the San Andreas fault system. Scholz (1990) predicts the thickness of the seismogenic layer, and hence the maximum focal depths of earthquakes, using both a similar temperature criterion as that described above and additional fault-velocity constraints.

Qualitative arguments of Sibson (1982) and Smith and Bruhn (1984) suggest that the theoretically derived transition depth from brittle to quasi-plastic flow for silica-rich rocks is controlled primarily by a critical temperature of approximately 350°C to 450°C and occurs at or near the depth of maximum shear stress (Figure 22). At this depth, short-term strain rates greater than 10^{-4}/sec are necessary to achieve brittle failure during earthquakes within the more ductile, intermediate-depth crustal material. In theory, this is the critical depth for nucleation of the largest magnitude earthquakes.

On the basis of the observed heat flow and extrapolated thermal gradients of the ISB, Smith and Bruhn (1984) inferred that this critical depth of earthquake nucleation would not exceed ~10 km, but large stress drops for magnitude-7 earthquakes could produce locally higher strain rates allowing their nucleation at mid-crustal depths of about 15 km ± 5 km. In the cooler lithosphere of the Colorado Plateau and Rocky Mountains east of the ISB, where background heat flow is less than 65 mW m^{-2}, maximum focal depths exceed 30 to 40 km and are attributed to deeper depths for the critical isotherms (Wong and Chapman, 1990).

The influence of shallow, high temperatures on earthquake depth distributions was described for the Yellowstone caldera (Fig. 17), which is characterized by an extremely high heat flow of 1500 mW m^{-2}. The observed lateral variation in focal-depth maxima reflects the combined influence of conductive and convective heat flow, hypothesized to produce the abrupt shallowing of the critical 350°C to 450°C isotherm beneath the inner caldera

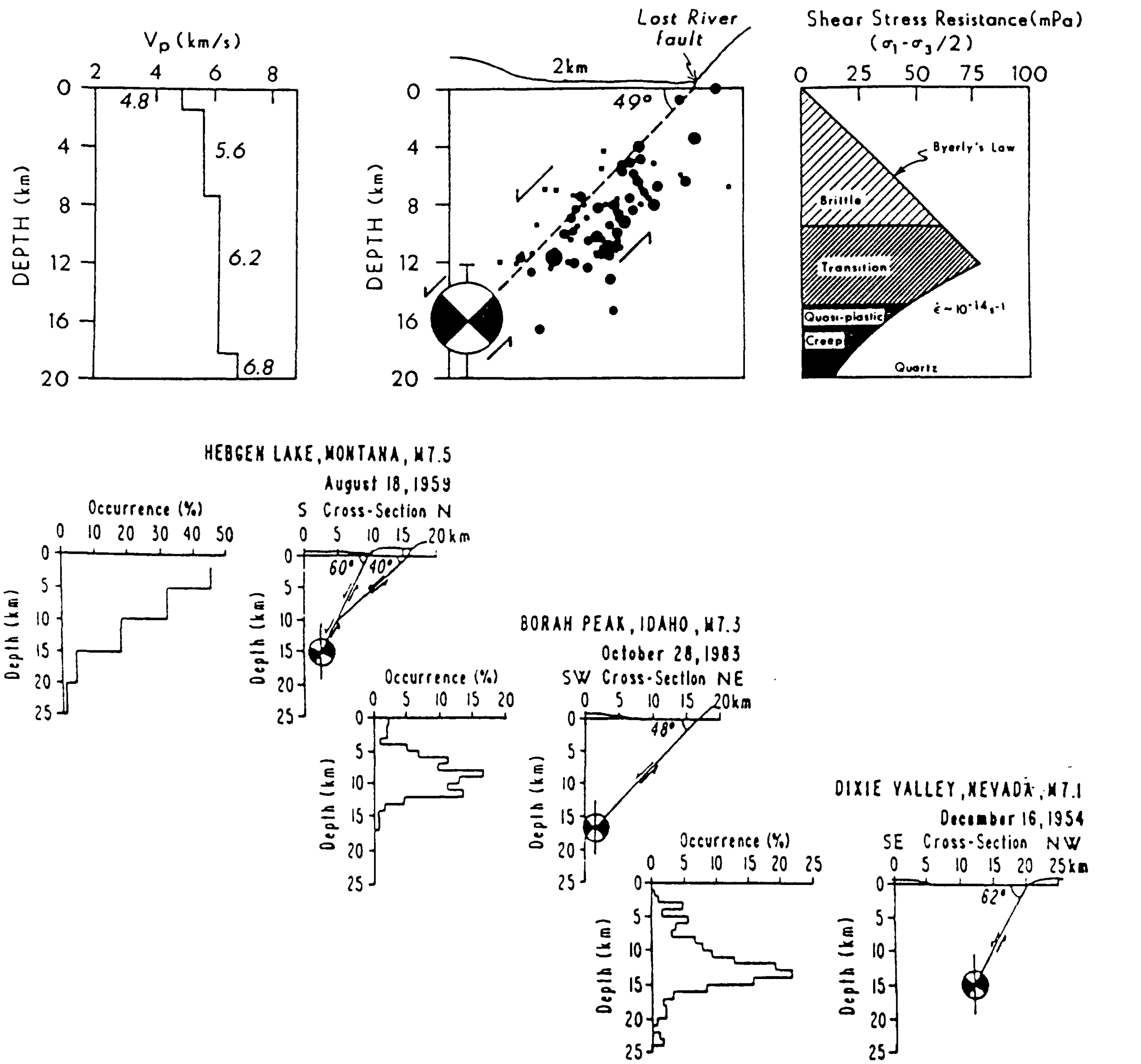

Figure 22. Hypothetical model for large (M >7.0) Basin and Range earthquakes, from Smith and others (1985). Upper, P-wave velocity model and subsurface fault geometry associated with the 1983, M_S = 7.3, Borah Peak, Idaho, earthquake (after Richins and others, 1985), together with a rheological model for the upper crust showing shear stress versus depth for a quartz rheology (after Smith and Bruhn, 1984). Lower, Fault-plane geometries and corresponding focal-depth histograms for three large historical earthquakes in the Basin and Range province; sources of data: Doser (1985b) and a compilation of microearthquake focal depths from University of Utah student theses, for the Hebgen Lake earthquake; Doser and Smith (1985) and Richins and others (1985), for the Borah Peak earthquake; and Okaya and Thompson (1985) and Doser (1986), for the Dixie Valley earthquake.

(Smith, 1989). Possibilities for the mechanisms controlling earthquake depth distributions include crystallization of rhyolite and basaltic melts, hydrothermal fluid flow into the shallow crust of the caldera, and the shallowing of superheated brines above a magma source (Pelton and Smith, 1982; Dzurisin and others, 1990).

Influence of the Yellowstone hot spot on the seismicity of the ISB

The occurrence of the 1983 Borah Peak earthquake on the flank of the SRP focused attention on the hypothesis that the Yellowstone hot spot and its track, the late Cenozoic Y-SRP volcanic province, are influencing the contemporary seismotectonics of the surrounding region (Smith and others, 1985; Scott and others, 1985; Anders and others, 1989). The unusual parabolic-shaped pattern of earthquakes surrounding the eastern SRP (Figs. 1 and 15) has been hypothesized to reflect the influence of lateral variations in deviatoric stresses, lithospheric subsidence, and high temperatures (Smith and others, 1985, in preparation; Blackwell, 1989), or an integrated loss of strength of upper-crustal volcanic material underlying the SRP (Anders and others, 1989). In all these hypothesized models, the thermomechanical response of the lithosphere may influence the seismogenic potential of the central ISB at map distances up to 100 km or more away from the SRP, thus altering the regional seismicity pattern.

The influence of the Yellowstone hot spot on the Quaternary tectonics of the ISB is further characterized in Fig. 23 (from Smith and others, 1990a), where late Cenozoic normal faults, calderas, and ages of plate and volcanic progression are plotted. A hypothesized "shoulder" of lithospheric subsidence flanking the SRP is also shown. Figure 23 shows that segments of normal faults with latest Quaternary displacements are systematically located away from the boundary of the SRP, whereas segments adjacent to the SRP are characterized by older Quaternary ruptures (Scott and others, 1985; Smith and others, 1985; Anders and others, 1989). Anders and others (1989) plot a double parabola with an apex at Yellowstone that envelops both the youngest faulting and background seismicity surrounding the relatively aseismic SRP. Their inner parabola follows approximately, but not exactly, the coarse dashed line sketched in Figure 23 marking the outer boundary of the subsidence shoulder around the SRP. The epicenter of the 1983 Borah Peak main shock is seen in Figure 23 to be on the outer boundary of the subsidence shoulder and, hence, along the inner margin of the SRP's flanking seismicity and along the boundary delimiting older and younger faulting.

Brott and others (1981) and Blackwell (1989) have shown that there is up to 700 m of systematic topographic subsidence southwestward along the SRP, away from the Yellowstone caldera along the 800-km-long track of the Yellowstone hot spot, that matches the cooling curve for the passage of a crustal heat source. These observations and the parabolic pattern of seismicity of the central ISB suggest that a thermal anomaly due to passage of the Yellowstone hot spot extends laterally beyond the SRP, systematically influencing regional seismicity in northern Utah, eastern Idaho, and western Wyoming.

Models for large Basin-and-Range-type normal-faulting earthquakes

In the past four decades, three of the largest earthquakes in the western United States were normal-faulting events that occurred in the Basin and Range province: the M_S = 6.8 Dixie Valley, Nevada, earthquake of 1954; the M_S = 7.5 Hebgen Lake, Montana, earthquake of 1959; and the M_S = 7.3 Borah Peak, Idaho, earthquake of 1983. Two of these earthquakes occurred in the ISB, and the other occurred in the central Basin and Range province. Studies of these earthquakes have provided new information on the geometry and mechanism of normal-faulting earthquakes, briefly summarized here. Some primary characteristics of these large earthquakes, illustrated in Figure 22, include rupture on planar normal faults, dipping 40° to 60°, and nucleation at mid-crustal depths of about 15 km, near the depth of the brittle/ductile transition.

Surface deformation for large normal-faulting earthquakes involves both coseismic deformation and slow deformation between the large earthquakes (King and others, 1988). Figure 24, from Smith and Richins (1984), shows a compilation of geodetic data representing basically coseismic ground deformation associated with the Dixie Valley, Hebgen Lake, and Borah Peak earthquakes. Hanging-wall subsidence dominates the observed surface deformation, which extends laterally up to 20 km from the surface fault trace and is a function of the causative fault's width, dip, and coseismic slip. For the Hebgen Lake earthquake, geodetic observations are not available for determining coseismic uplift of the footwall. To model the surface deformation of large normal-faulting earthquakes, King and others (1988) calculated the coseismic and long-term response due to a planar shear-dislocation within an elastic layer underlain by a viscous half-space. Their results show that, for coseismic deformation, hanging-wall subsidence is predicted to be roughly two to four times greater than footwall uplift for faults dipping 45° to 60°; long-term adjustments involving stress relaxation and deformation due to erosion and sediment-deposition tend to broaden the profile of surface deformation, thereby slightly raising both the hanging-wall and footwall blocks (King and others, 1988).

Low-angle and listric faults

The working model outlined in Figure 22 and described above for large normal-faulting earthquakes is not simply compatible with observations from seismic-reflection data and geologic mapping that document Quaternary listric and low-angle normal faulting in many subsurface locations in the central and eastern Basin and Range province (Royse and others, 1975; Anderson and others, 1983; Smith and Bruhn, 1984; Smith and others, 1989). These structures typically flatten at maximum depths of 4 to 6 km, at similar depths to much of the Sevier Desert detachment of western Utah, described in the section on seismotectonic framework.

The question of the seismogenic capability of shallow dipping faults in the ISB has been addressed by various workers by scrutinizing compilations of focal mechanisms (e.g., Zoback, 1983; Arabasz and Julander, 1986; Bjarnason and Pechmann, 1989; Doser and Smith, 1989), and all have similarly found the predominance of seismic slip on planes of moderate (> 30°) to steep dip, with mean dips in the range of 45° to 60°. Some fault-plane solutions can be found with one low-angle nodal plane; however, in those cases there is no corroborating evidence in the form of clustered earthquake foci on either a downward-flattening or planar low-angle normal fault to support selection of the low-angle nodal plane as the plane of seismic slip (Arabasz

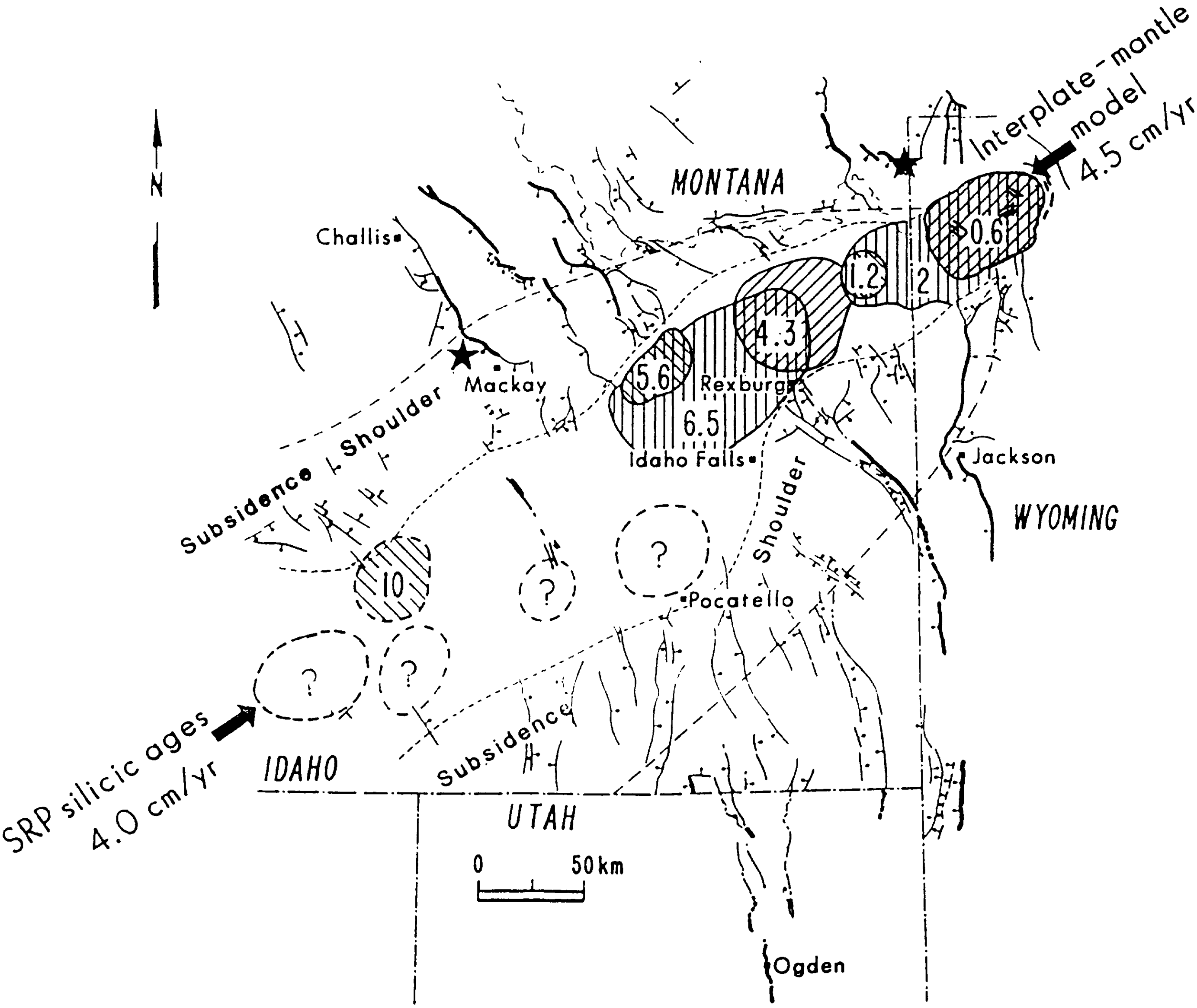

Figure 23. Seismotectonic framework of the Snake River Plain–Yellowstone region, from Smith and others (1990a). Map shows the northeastward decreasing age, in millions of years, of volcanic calderas (hachured) along the eastern Snake River Plain (ESRP), associated with passage of the Yellowstone hot spot, together with the boundary of an hypothesized shoulder of lithospheric subsidence (coarse dashed line), and late Quaternary normal faulting (marked by a heavy line weight where most recent displacement is Holocene). Arrow at Yellowstone (upper right) shows the direction of motion of the North American plate over the Yellowstone hot spot at a relative velocity of 4.5 cm/yr—in agreement with the northeastward space-time progression of calderas at a rate of 4.0 cm/yr. Fine dashed line indicates boundary of the Snake River Plain volcanic province; stars, the epicenters of the 1959 Hebgen Lake and 1983 Borah Peak earthquakes, as in Figure 9.

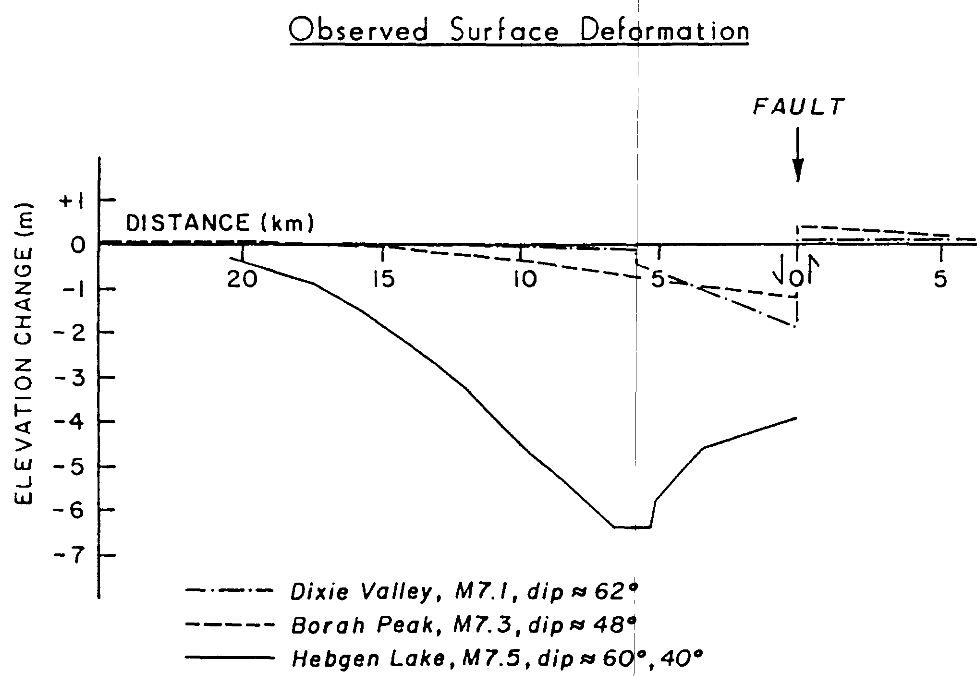

Figure 24. Coseismic vertical ground deformation, chiefly hanging-wall subsidence, produced by the three large scarp-forming, normal-faulting earthquakes identified in Figure 22, from Smith and Richins (1984). Sources of data: Savage and Hastie (1969) and Reil (1957), for the Dixie Valley earthquake; Stein and Barrientos (1985), for the Borah Peak earthquake; and Savage and Hastie (1966), for the Hebgen Lake earthquake.

and Julander, 1986). An investigation by Doser and Smith (1989) of 57 focal mechanisms for moderate to large normal and oblique-slip earthquakes throughout the U.S. Cordillera, including the ISB, revealed no evidence for nodal planes shallower than 30°. Jackson and White (1989) examined a global distribution of normal-faulting focal mechanisms with the same finding. Thus there appears to be no present evidence for the nucleation or slip of an historical earthquake in the ISB on a low-angle or listric normal fault.

Satisfactory answers to questions about the mechanical origin and seismic capability of low-angle normal faults are elusive. Melosh (1990) recently posed a mechanism for listric faulting in the Basin and Range that required a rotation of the maximum and minimum principal stresses, with the principal stress vertical at the surface and rotating to 45° at depths of a few kilometers. His model assumes an elastic upper crust and a viscoelastic lower crust, where the principal-stress rotation flattens the fault. Although this is an attractive hypothesis, the occurrence of large normal-faulting earthquakes at mid-crustal depths of about 15 km on planar faults dipping 45° to 60° argues for a homogeneous stress field to at least the maximum depth of crustal earthquakes in the Intermountain region.

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Appendix D

Document (Machette and others, 1991) presented to NEPEC by Machette, summarizing what is known about the Wasatch fault zone

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The Wasatch fault zone, Utah—segmentation and history of Holocene earthquakes

MICHAEL N. MACHETTE, STEPHEN F. PERSONIUS and ALAN R. NELSON

U. S. Geological Survey, Denver, Colorado, U.S.A.

DAVID P. SCHWARTZ

U.S. Geological Survey, Menlo Park, California, U.S.A.

and

WILLIAM R. LUND

Utah Geological and Mineral Survey, Salt Lake City, Utah, U.S.A.

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Abstract—The Wasatch fault zone (WFZ) forms the eastern boundary of the Basin and Range province and is the longest continuous, active normal fault (343 km) in the United States. It underlies an urban corridor of 1.6 million people (80% of Utah's population) representing the largest earthquake risk in the interior of the western United States.

We have used paleoseismological data to identify 10 discrete segments of the WFZ. Five are active, medial segments with Holocene slip rates of 1–2 mm a^{-1}, recurrence intervals of 2000–4000 years and average lengths of about 50 km. Five are less active, distal segments with mostly pre-Holocene surface ruptures, late Quaternary slip rates of <0.5 mm a^{-1}, recurrence intervals of ≥10,000 years and average lengths of about 20 km. Surface-faulting events on each of the medial segments of the WFZ formed 2–4-m-high scarps repeatedly during the Holocene; latest Pleistocene (14–15 ka) deposits commonly have scarps as much as 15–20 m in height. Segments identified from paleoseismological studies of other major late Quaternary normal faults in the northern Basin and Range province are 20–25 km long, or about half of that proposed for the medial segments of the WFZ.

Paleoseismological records for the past 6000 years indicate that a major surface-rupturing earthquake has occurred along one of the medial segments about every 395 ± 60 years. However, between about 400 and 1500 years ago, the WFZ experienced six major surface-rupturing events, an average of one event every 220 years, or about twice as often as expected from the 6000-year record. This pattern of temporal clustering is similar to that of the central Nevada–eastern California Seismic Belt in the western part of the Basin and Range province, where 11 earthquakes of $M > 6.5$ have occurred since 1860. Although the time scale of the clustering is different—130 years vs 1100 years—we consider the central Nevada–eastern California Seismic Belt to be a historic analog for movement on the WFZ during the past 1500 years.

We have found no evidence that surface-rupturing events occurred on the WFZ during the past 400 years, a time period which is twice the average intracluster recurrence interval and equal to the average Holocene recurrence interval. In particular, the Brigham City segment (the northernmost medial segment) has not ruptured in the past 3600 years—a period that is about three times longer than this segment's average recurrence interval during the early and middle Holocene. Although the WFZ's seismological record is one of relative quiescence, a comparison with other historic surface-rupturing earthquakes in the region suggests that earthquakes having moment magnitudes of 7.1–7.4 (or surface-wave magnitudes of 7.5–7.7)—each associated with tens of kilometers of surface rupture and several meters of normal dip slip—have occurred about every four centuries during the Holocene and should be expected in the future.

INTRODUCTION

SEVERAL problems are central to understanding the processes and timing of earthquakes associated with surface-faulting events in extensional terrains: (1) the identification of long (>50-km) faults; (2) the determination of the occurrence and nature of characteristic earthquake events; and (3) the determination of the length and variability of recurrence intervals for large magnitude earthquakes, which could be potentially devastating to the region. We have gained greater insight into these problems through paleoseismological investigations involving detailed geologic mapping and exploratory trenching across the Wasatch fault zone (WFZ) in Utah.

Regional setting

The WFZ is one of the longest and most active extensional fault zones in the western United States. Its late Quaternary trace extends for 343 km from Malad City, Idaho, to Fayette, Utah, and is marked by large (30–50-m-high) scarps on glacial, lacustrine, colluvial and alluvial deposits of middle to late Pleistocene age and smaller (3–10-m-high) scarps on Holocene deposits. The WFZ is the main component of a prominent struc-

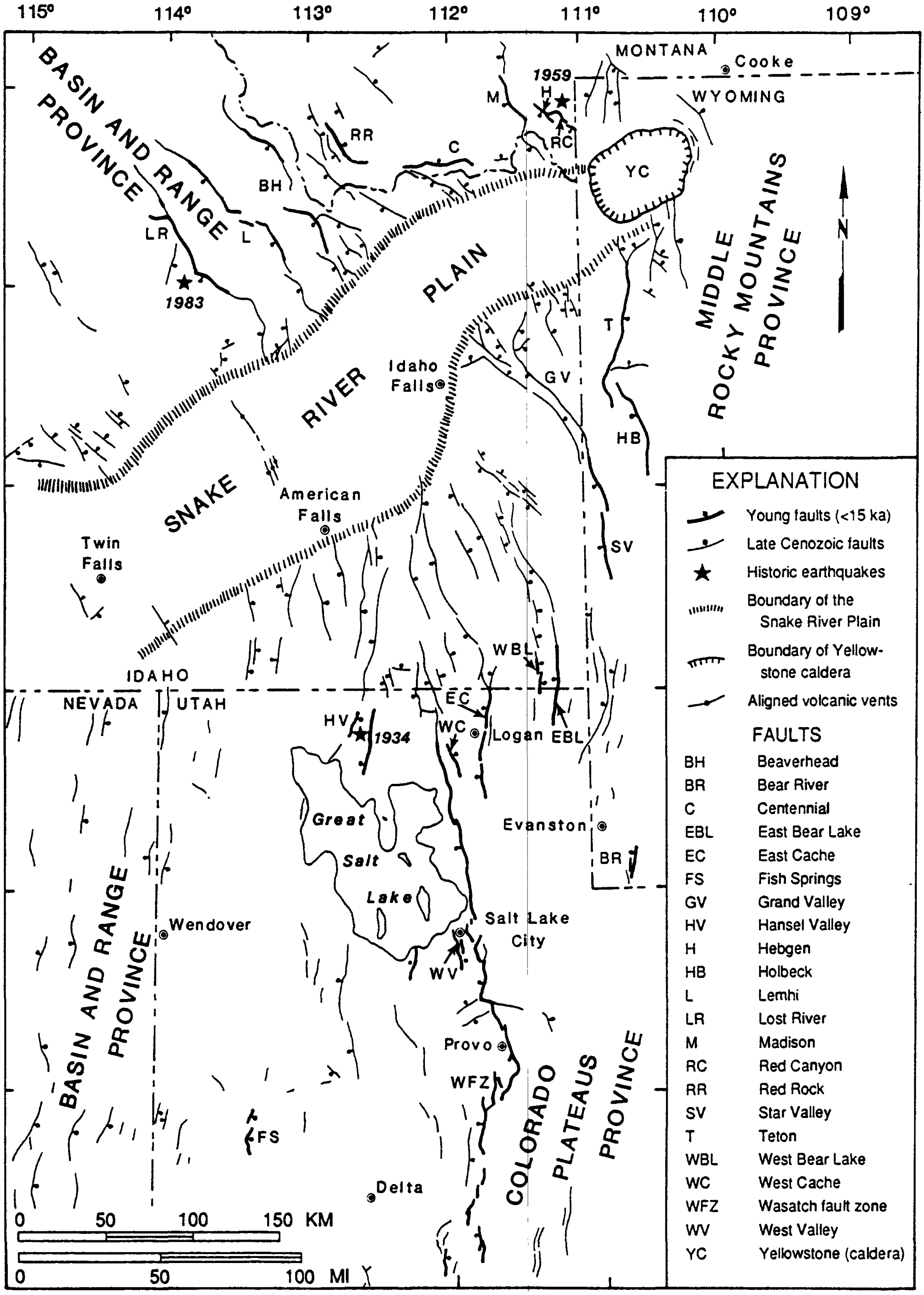

Fig. 1. Index map showing selected major late Cenozoic faults in the northeastern part of the Basin and Range province and the northern part of the Intermountain Seismic Belt. Faults having young movement (<15 ka) are shown by bold lines. (Compiled from maps of Howard *et al.* 1977, Nakata *et al.* 1982, Anders *et al.* 1989, Machette *et al.* in press. Smith *et al.* in press.)

tural transition zone that separates the greatly extended terrain of the Basin and Range province from the uplifted Colorado Plateaus and Middle Rocky Mountains provinces (Fig. 1). Concentrated seismicity along this zone (the Intermountain Seismic Belt, Smith & Sbar 1974) is largely coincident with a belt of young faulting (<15 ka) that forms a right-stepping en échelon pattern from the northern part of the WFZ to the Yellowstone area; from there, it trends westward across southwestern Montana into central Idaho and includes the two largest historic earthquakes in the region (Fig. 1). This V-shaped pattern of recent tectonism flares out from the Snake River Plain and the tip of the V is centered on the Yellowstone calderas, which are the present location of the Yellowstone hotspot (see Scott *et al.* 1985, Smith *et al.* 1985, Pierce & Scott 1986, Smith 1988, Anders *et al.* 1989, Pierce & Morgan 1990, Smith *et al.* in press). The Yellowstone hotspot forms a regional thermal anomaly (Quaternary volcanism, mantle upwelling and high heat flow), which has migrated to the northeast along the Snake River Plain during the past 15 Ma at an average rate of 3–4 cm a^{-1} (see discussions in Pierce & Scott 1986, Pierce & Morgan 1990). Although these authors discuss various hypotheses about the relation between the Yellowstone hotspot and its effect on late Cenozoic tectonism, all agree that the track of the Yellowstone hotspot has greatly influenced the region's temporal and spatial pattern of faulting in the past, and may do so in the future.

Previous studies

Although the WFZ has been the subject of scientific interest since the pioneering work of G. K. Gilbert a century ago (Gilbert 1890, Machette 1988a), the first comprehensive study was undertaken in the 1970s by Cluff *et al.* (1970, 1973, 1974) using low-sun-angle aerial photographs to map the surface trace of the WFZ. As an extension of their reconnaissance mapping, geologists at Woodward-Clyde Consultants made detailed investigations at four trench sites along the WFZ during the period 1978–1982. Their work culminated in two major synthesis reports that applied new concepts to the paleoseismological history of the WFZ. In the first report, Swan *et al.* (1980) proposed segmentation of the WFZ and speculated on the number of possible segments; they suggested at least six on the basis of modern microseismicity to as many as 10 on the basis of geometric variations along the fault zone and the commonly observed rupture length of 30–40 km for historic normal faults. In the second report, Schwartz & Coppersmith (1984) proposed that the WFZ is composed of six major segments that were chosen on the basis of a combination of geomorphic, topographic, geophysical, paleoseismic and geodetic data.

The concept of fault segmentation embodies the idea that major slip events (long surface ruptures having several meters of offset that are associated with large-magnitude earthquakes) on normal fault zones are largely confined to discrete parts that represent only a fraction of the fault's total length and whose boundaries are related to geometric and structural controls along the fault zone. However, the term *segment* was not explicitly defined in either of these reports; recently, the term has been used in various contexts that range from "a portion" (e.g. a geometric segment) to "a structural entity" (e.g. a structural segment). dePolo *et al.* (1991) discussed this nomenclature problem and suggested that the term "earthquake segment" be used for those parts of a fault or faults that "rupture as a unit during an earthquake". In this paper, we use the term *segment* in the same way, although our determination of segments of the WFZ is based on paleoseismology rather than contemporary seismology. The segments defined herein indicate the extent of surface rupturing that we would expect during large-magnitude earthquakes that nucleate on the WFZ.

Recent studies

Our investigations of the WFZ during the 1980s were conducted under the U.S. Geological Survey's (USGS) National Earthquake Hazards Reduction Program, which focuses on analysis of seismic risk in populated regions with earthquake hazards. The first phase was to make 1:50,000-scale surficial geologic maps of the Wasatch Front region (Personius 1988, 1990, Machette 1989, in press, Nelson & Personius 1990, in press, Personius & Scott 1990, in press) as a basis for derivative studies (potential shaking, liquefaction, hazard assessments, special study zones, etc.). The results of initial mapping of the WFZ led Machette *et al.* (1986) to modify some of Schwartz & Coppersmith's (1984) proposed segment boundaries, suggested several new boundaries and subdivided four of the original segments. These modifications were based on recent fault movements as determined by analysis of fault-scarp morphology and from the relations between young surficial deposits and fault scarps. In 1986, the USGS and the Utah Geological and Mineral Survey initiated a second phase with a co-operative program of exploratory trenching to test segmentation models by determining (1) the recency of faulting along several of the newly proposed segments, (2) the timing and recurrence intervals of older faulting events and (3) the timing of movement on adjacent fault segments (i.e. presence or lack of synchronous movement). Trenching studies have been completed at sites near Brigham City, at East Ogden, Dry Creek (south of Salt Lake City), at the mouths of American Fork and Rock Canyons, and near Mapleton (see Machette *et al.* in press, fig. 1 and appendix). In addition, co-operative studies with geologists at the Bureau of Reclamation, University of Colorado, and Utah State University have augmented our own findings. This recent flurry of co-operative investigations has more than tripled the number of trenches available in 1980, increased our data on the timing of faulting events by a factor of five, and significantly tightened the error limits on previously documented faulting events.

This paper is our attempt to reach a consensus on a segmentation model and the movement history for the WFZ based on our collective studies. However, as work continues on the WFZ and as new sites are explored, we anticipate further refinements in our segmentation model and timing history.

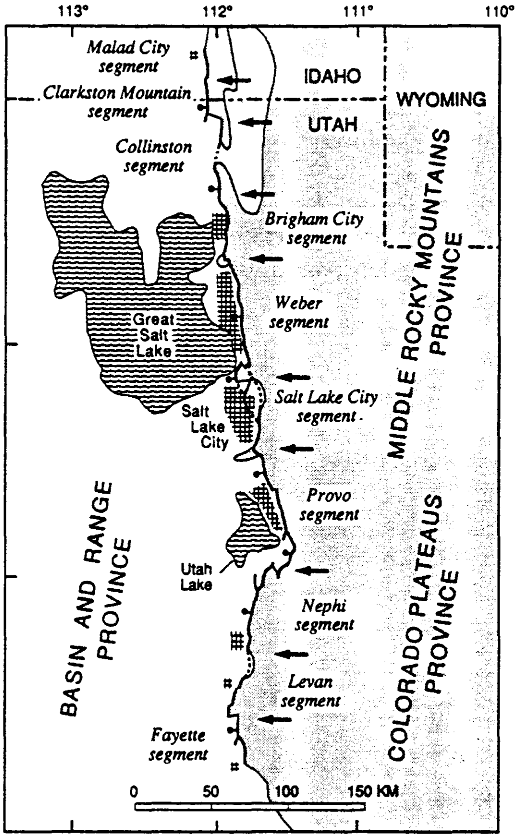

Fig. 2. Location of segments of Wasatch fault zone. Solid arrows indicate segment boundaries. Major towns shown by cross-hachure symbol.

SEGMENTATION OF THE WASATCH FAULT ZONE

The central two-thirds (Brigham City to Nephi segments) of the WFZ has ruptured two or more times in the past 6000 years (Fig. 2). The most recent movement on the four other distal segments is considered to be pre-Holocene because we have not found fault scarps on uppermost Quaternary deposits along these segments. The distal portions have short segments, low slip rates (<0.5 mm a^{-1}), and long recurrence intervals (>10,000 years) whereas the medial segments along the central portion of the WFZ are characterized by long segments, high slip rates (1–2 mm a^{-1}) and recurrence intervals of about 2000 years. Obviously, the WFZ is a fundamental province-scale fault zone that serves to decouple the greatly extended terrain of the Basin and Range province from the more intact provinces to the east (Fig. 1).

Segmentation models

The six segments originally proposed by Schwartz & Coppersmith (1984) are retained in our current model with some changes in position and nomenclature. Although they ended their northernmost (Collinston) segment at the Bear River, they followed the belief of Cluff *et al.* (1970) that older movement on the WFZ continued farther to the north into southern Idaho (Fig. 2). The southern end of their Collinston segment has been relocated 5 km north of Brigham City by Personius (1988) and its northern end has been extended about 11 km past the Bear River (Machette *et al.* 1987, in press). Machette *et al.* (1987) also extended the Wasatch fault zone north along the Malad Range to Malad City, Idaho. Although no detailed paleoseismic studies have been conducted on this portion of the WFZ, the northern 41 km has been subdivided tentatively into the Clarkston Mountain and Malad City segments (Fig. 2), which are separated by a 6-km-wide salient. The original Ogden segment of Schwartz & Coppersmith (1984), which extended from north of Brigham City to northern Salt Lake City, is now divided into the 40-km-long Brigham City segment and the 61-km-long Weber segment. The Salt Lake City segment remains virtually unchanged from Schwartz & Coppersmith's proposed location and boundaries.

The Provo segment, which borders the eastern margin of Utah Valley (Machette 1989, in press) was named by Schwartz & Coppersmith (1984), but Machette *et al.* (1986) subdivided this part of the WFZ into three shorter segments (American Fork, Provo—restricted sense, and Spanish Fork) on the basis of apparent recency of movement as determined from scarp morphology and detailed mapping of the fault zone. However, trenching at three new sites has lead us to conclude that the entire length (70 km) of the range-bounding WFZ in Utah Valley is a single segment (Provo). This conclusion is based on similarities in the timing of the most recent (500–650 years ago) and penultimate (2.6–3.0 ka) events determined from trenching at the American Fork Canyon (Machette 1988b, Forman *et al.* 1989) and Mapleton sites (Schwartz *et al.* 1988, Lund *et al.* in press). The most recent movement along the central part of the segment had been poorly constrained at about 1100 years ago (Machette *et al.* 1987), but recent trenching at the Rock Canyon site constrains it to between 600 and 2700 years B.P., which is compatible with events at the American Fork and Mapleton sites (Lund *et al.* in press).

The Nephi segment (Fig. 2) is the southernmost segment of the WFZ that shows demonstrable evidence of repeated Holocene movement and is one of the most recently active segments. A 15-km-long gap in latest Quaternary faulting separates the Nephi segment from the Levan and Fayette segments, which comprise the southern part of the WFZ. Machette's studies of fault-scarp morphology suggest that scarps along the Levan segment are distinctly younger than those along the Fayette segment, and thus comprise two discrete fault segments. The Fayette scarps are probably early Holocene(?) or latest Pleisocene in age, whereas the Levan scarps are latest Holocene in age. The southern part of the WFZ is characterized by low slip rates (<0.5 mm a^{-1}) and recurrence intervals that are >6000 years (Levan segment) to ≤10.000 years (Fayette segment).

Segment lengths

The total length of the WFZ is about 383 km as measured along its surface trace, which is our preferred method of reporting fault lengths, and about 343 km from end-to-end (Table 1). However, the net length of surface trace does not include overlapping portions or gaps in the cumulative lengths. Individual segments are as short as 11–17 km on the distal parts to as long as 60 km on the Weber segment and about 70 km on the Provo segment. The average length along surface trace for all 10 segments of the WFZ is about 37 ± 19 km. The five medial segments average about 52 ± 13 km in length along their surface trace. The five distal segments (three on the north and two on the south) average about 21 ± 8 km in length, with the inboard segments (Levan and Collinston) each being of intermediate length (30 km). In general, the distribution of both segment lengths and slip rates on the WFZ forms a broad envelope with maximum values in the central part and decreasing values at the ends, as reflected by the altitude of the crest of the Wasatch and associated ranges along the fault zone (see Schwartz & Coppersmith 1984, fig. 10). In general, these relations indicate a strong positive correlation between topographic relief along the WFZ (a proxy for structural offset), slip rates and segment lengths, and an inverse correlation between topography and recurrence intervals of major surface-faulting events.

Segment boundaries

Many normal fault zones present continuous structural pathways for surface rupturing that are one hundred to several hundred kilometres long. Typically, individual surface-faulting events can rupture as much as 50 km of ground—only a fraction of the total length of fault zones. Studies of both historic and prehistoric faulting show that rupturing tends to occur in discrete sections (or segments) of fault zones that are separated by boundaries.

If boundaries between segments of long fault zones are persistent barriers to lateral propagation of earthquake ruptures, then they should coincide with structural anomalies along the fault. These anomalies may be coincident with abrupt changes in structural relief along strike (Wheeler 1989) or with areas of increased fault complexity (King 1986, Bruhn *et al.* 1987). Changes in structural relief may be expressed as: (1) subsurface

Table 1. Lengths of Wasatch fault zone segments and positions of boundaries (lengths are rounded to closest 0.5 km)

Fault segment	Length (km)		Comments
	Surface trace	Straight line	
Malad City	17.0	16.5	Last movement >14 ka. Does not include 6-km Woodruff spur to south
Clarkston Mountain	19.0	17.0	Last movement >14 ka. Extends from Woodruff spur to Malad River. Has 7-km left-step and 2-km overlap with Collinston segment
Collinston	30.0	29.5	Last movement >14 ka. North end at Short Divide; position of fault adjacent to Bear River uncertain
Brigham City	40.0	35.5	Repeated Holocene movement. Has 1-km left step and 1.5-km overlap with Weber segment at south end
Weber	61.0	56.0	Repeated Holocene movement. South end on north-central flank of Salt Lake salient. Steps 2.7-km west to Warm Springs fault
Salt Lake City	46.0	39.0	Repeated Holocene movement. Has three left-stepping surface traces: Warm Springs (10 km), East Bench (13 km) and main Wasatch (23 km; Cottonwood section). South end steps 7.5 km east across Traverse Range on Fort Canyon fault
Provo	69.5	59.0	Repeated Holocene movement. Extends from Traverse Range to Payson Canyon. Has overlap and right step to Nephi segment
Nephi	42.5	37.5	Repeated Holocene movement. Extends from Payson to Nephi, steps 8.5-km to Juab Valley. Separated from Levan segment by 15-km gap
Levan	30.0	25.5	One Holocene movement. Includes two major gaps (6-km net) within segment. Steps 3.5 km east and 5 km south to Fayette segment
Fayette	11.0	10.5	No Holocene movement. Has 4-km-long western strand and 9-km-long range-bounding eastern strand
Entire WFZ	383.0	343.0	Total length of all segments from end to end
All segments	366.0	326.0	Net length of all segments of the WFZ, excluding gaps
Average segment	36.6 ± 19.0	32.6 ± 16.2	Length of all segments divided by 10
Holocene segments	259.0	227.0	Sum of lengths of segments with repeated Holocene movement
Average Holocene segment	51.8 ± 12.8	45.4 ± 11.1	Length of Holocene segments divided by 5
Older segments	107.0	99.0	Sum of lengths of segments without repeated Holocene movement
Average older segment	21.4 ± 8.4	19.8 ± 6.8	Length of older segments divided by 5

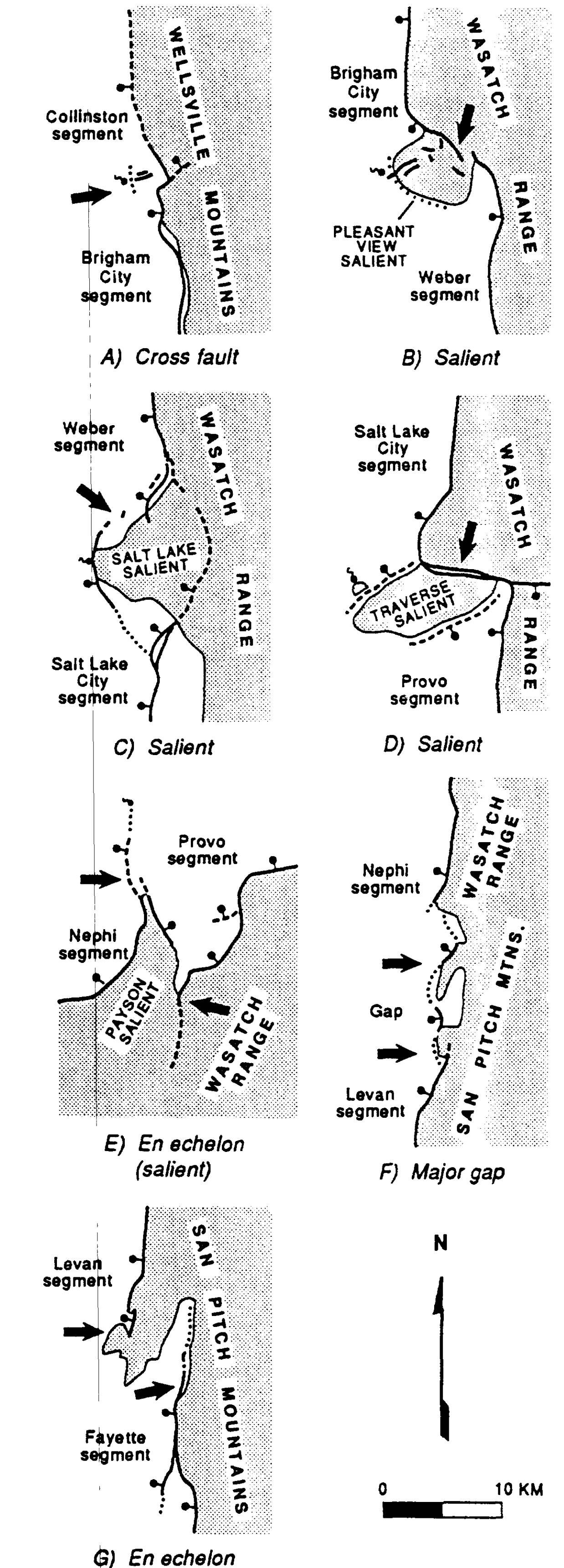

Fig. 3. Examples of map patterns at segment boundaries along the Wasatch fault zone in northern Utah. Bedrock shown by stipple pattern. Figures are listed from north (A) to south (G).

bedrock ridges between deep structural basins (i.e. gravity saddles, Zoback 1983); (2) depressed structural levels in the footwall; (3) bedrock blocks (e.g. salients) stranded at intermediate structural levels between parallel strands of the fault zone; or (4) a combination of these features. Our studies show that structural boundaries along the WFZ probably have terminated or arrested the propagation of earthquake ruptures repeatedly in the Holocene (see next section).

Most persistent segment boundaries on the WFZ are associated with major bedrock blocks (salients or spurs) that extend into the basin at intermediate structural levels (Fig. 3). These salients commonly are bounded by Quaternary faults that are less active than the range-bounding faults. The bedrock spurs south of Malad City, and north of Ogden (Pleasant View, Fig. 3b) and Salt Lake City (Fig. 3c) and the Traverse Range (Fig. 3d) are salients. The Traverse Range is detached from the Wasatch Mountains along the Fort Canyon fault, which is the westward extension of the S-dipping low-angle Deer Creek fault (Bruhn *et al.* 1987). Salients can persist for millions of years as barriers to laterally propagating ruptures along a fault zone. Some persistent barriers, however, may not fully arrest the propagation of rupturing. For example, Ostenaa (1990) argues that the Provo–Nephi boundary (Fig. 3e) may be a "leaky" barrier (terminology of Crone & Haller 1991); that is, one which allows partial or sympathetic rupture of an adjacent segment as occurred in the 1983 rupture of the Warm Springs and Thousand Springs segments of the Lost River fault zone (Crone *et al.* 1987).

Several of the segment boundaries along the WFZ are associated with major en échelon (lateral) steps. These steps typically cross bedrock-cored ranges, such as at Dry Mountain (Fig. 3e, Payson salient) and between the Levan and Fayette segments (Fig. 3g), and can be considered as a variety of salient bounded by active faults on both sides. Most steps that are not associated with salients are non-persistent barriers; e.g. dePolo *et al.* (1991) and Zhang *et al.* (1991) showed that during the 1954 Fairview Peak, Nevada, earthquake surface faulting extended across several en échelon steps and, thus, failed to arrest or stop the lateral propagation of ruptures.

Cross faults or other intersecting faults also occur at persistent boundaries. This type of boundary involves oblique intersection of two or more fault traces. The cross faults typically extend into bedrock (e.g. the northern part of the Brigham City segment) (Fig. 3a). The boundary between the Collinston and Clarkston Mountain segments is formed by an E–W cross fault that links two en échelon segments. These types of fault intersections are not usually associated with reduced structural relief in either the ranges or basin and, thus, may be largely non-persistent.

Several other morphological features are present along fault segments that do not appear to be unique criteria for differentiating segment boundaries. The most common of these are geometric changes in fault zones, such as changes in fault strike, branching faults (bifurcations), some en échelon steps, and gaps in faulting.

Abrupt bends in fault traces, such as the 110° bend on the Provo segment at Spanish Fork Canyon, rarely form segment boundaries or persistent barriers to rupturing. Crone & Haller (1991) found that concave bends in the down-dip directions of late Quaternary faults north of the Snake River Plain (Fig. 1) typically do not form segment boundaries, whereas convex bends may be persistent boundaries. Bruhn *et al.* (1987) suggest that bifurcations in fault zones may be areas of increased fault roughness and thus points of rupture nucleation or termination. Bruhn *et al.* (1987, table 1) identified five potential barriers to rupturing along the Salt Lake City segment of the WFZ by analyzing paleoslip directions, stress tensors, and fault-zone rupture characteristics. They suggest that non-conservative barriers at the ends of the segments (the Salt Lake Salient and the Transverse Mountains, Figs. 3c & d, respectively) and at the bifurcation zone between the southern part of the Salt Lake segment (Cottonwood section, Table 1) and the East Bench fault (Table 1) may control the propagation of ruptures during a large earthquake. Although this hypothesis cannot be tested because of a lack of good trenching sites, Personius & Scott (1990, in press) argue that the bifurcation zone is probably a less resistant barrier that has directed rupturing from the Cottonwood section away from the range front and onto the East Bench fault, thereby leaving the range-bounding fault inactive along the northern part of the segment. A similar bifurcation zone at the Springville fault on the Provo segment (Table 1) appears to be a non-persistent barrier to rupture propagation (Machette in press).

Most of the gaps in surface rupturing along the WFZ are the result of en échelon steps that are several kilometers wide. However, a 15-km-long gap in recent faulting south of Nephi (Fig. 3f) marks the boundary between the Nephi and Levan segments (Machette *et al.* in press). This boundary appears to be a barrier that has persisted for tens of thousands of years. Holocene and uppermost Pleistocene alluvium (<15 ka) is not faulted in the gap, whereas upper and middle(?) Pleistocene alluvium has been offset tens of meters. The older scarps in the gap attest to the presence of a through-going WFZ that now is inactive. Long gaps such as this may persist for hundreds of thousands of years, but do not appear to result in anomalous structural relief along the fault zone.

Patterns of en échelon steps, bifurcations and gaps are examples of the types of geometric boundaries along fault zones that rarely are associated with persistent rupture barriers. In this context, some geometrically defined segments (such as the three proposed by Machette *et al.* 1986 for the Utah Valley part of the WFZ) are not persistent rupture entities but, rather, may be building blocks for long ruptures associated with large-magnitude earthquakes.

TIMING AND RECURRENCE OF HOLOCENE MOVEMENT ON THE WASATCH FAULT ZONE

Since the late 1970s, extensive efforts have been made to date individual earthquake events along the WFZ through trenching of productive sites. As of 1987, more than 45 trenches and natural exposures had been logged and described on six segments of the WFZ (Machette *et al.* 1987, table 1); since then, several additional sites have been investigated. Most of the trench sites have provided some control on the time of most recent faulting and set limits on recurrence intervals and slip rates. Since beginning our co-operative effort in 1985, we have obtained about 50 radiocarbon dates using both conventional and accelerator-mass-spectrometry methods on charcoal and soil organic matter and 17 experimental thermoluminescence age estimates. These ages (see appendix in Machette *et al.* in press) have been used to construct a chronology of Holocene surface-faulting and recurrence intervals for the WFZ. More importantly, we have used this chronology as the primary tool for defining segments along the zone.

Recurrence intervals

Our calculations of the average recurrence interval for segments having repeated Holocene movement are shown in Table 2. The *average* recurrence interval (*RI*) on any single segment is about 1980 ± 310 years. However, there is so much variation between and within segments that this value has little meaning. The *composite* recurrence interval (*CRI*), which is defined as the average time between two faulting events anywhere on the central part of the WFZ is 395 ± 60 years. Schwartz & Coppersmith (1984) reported a maximum recurrence interval (*CRI*) of 615–666 years, but preferred a value of 444 years, which is within the error limits of our new value. Even though the two methods of calculation were somewhat different, both investigations reached basically the same conclusion—i.e. a major surface-rupturing earthquake has struck the Wasatch Front once every four centuries (on average) during the past 6000 years.

Figure 4 shows our chronology of faulting along the medial segments of the WFZ. The recurrence intervals on these segments may vary from as little as 500 years (for the past two events on the Weber segment) to as much as 4000 years (on the Salt Lake City segment). At least one segment (Provo) has had two recurrence intervals of similar duration. Of particular interest is the lack of movement along the Brigham City segment during the past 3600 years. Two faulting events occurred at about 3.6 and 4.7 ka on the Brigham City segment, and a third between 4.7 and 6–8 ka (see Machette *et al.* 1987, in press)—an average of one event every 1500–2200 years. However, the two most-recent faulting events yield a recurrence interval of only 1100 years (Fig. 4). Of the medial segments, only the Brigham City has been inactive longer than its average recurrence interval. Thus, although the seismological record for the WFZ is one of relative quiescence, the paleoseismological record suggests that a major earthquake associated with tens of kilometers of surface rupture and several meters of normal dip-slip should be expected in the future.

Table 2. Timing, number of major surface-faulting earthquakes, and recurrence intervals for Holocene movement of the Wasatch fault zone

Fault segment	Trench site	(A) Oldest event (t)* or datum (d)† (years ago)	(B) Estimated time of MRE (years ago)	(C) Time interval (A − B, years)	(D) Number of events (E) and intervals (I): E	I
Brigham City	Brigham City	4700 ± 500t	3600 ± 500	1100 ± 1000	2	1
Weber	East Ogden	3750 ± 250t	500 ± 300	3250 ± 550	4	3
Salt Lake City	Dry Creek	5250 ± 250t	1500 ± 300	3750 ± 550	2	1
Provo	American Fork	5300 ± 300t	500 ± 200	4800 ± 500	3	2
Nephi	North Creek	5500 ± 200d	≥400	4900 ± 200	3	2
Levan	Deep Creek	7300d	1000	N/A	1	0
Totals (based on segments 1–5)				17,800 ± 2800	15	10
Calculated recurrence intervals for WFZ segments that have repeated Holocene movement‡				Value and error limit (years)		
Average recurrence interval (*RI*)				1980 ± 310		
Average composite recurrence interval (*CRI*; *RI*/5)				395 ± 60		

Note: All values for age and time intervals (columns A–C) are rounded to the nearest 100 years. Ages based on calendar-corrected radiocarbon dates and thermoluminescence analyses. The average recurrence interval is determined by dividing the sum of time intervals (column C) by the sum of intervals between faulting events (column D). Time interval (column C) for Nephi segment includes time between the oldest (undated) event at site and the age of the datum; thus, value in column (C) is a maximum. MRE, most recent faulting event; N/A, not applicable.

*t—time of oldest well-dated faulting event (rounded to nearest 50 years).

†d—age of datum from dating, stratigraphic, or tectonic considerations (rounded to nearest 50 years).

‡Three significant figures are used to compute average values of recurrence from the totals in columns (C) and (D). Values are rounded to nearest 5 years.

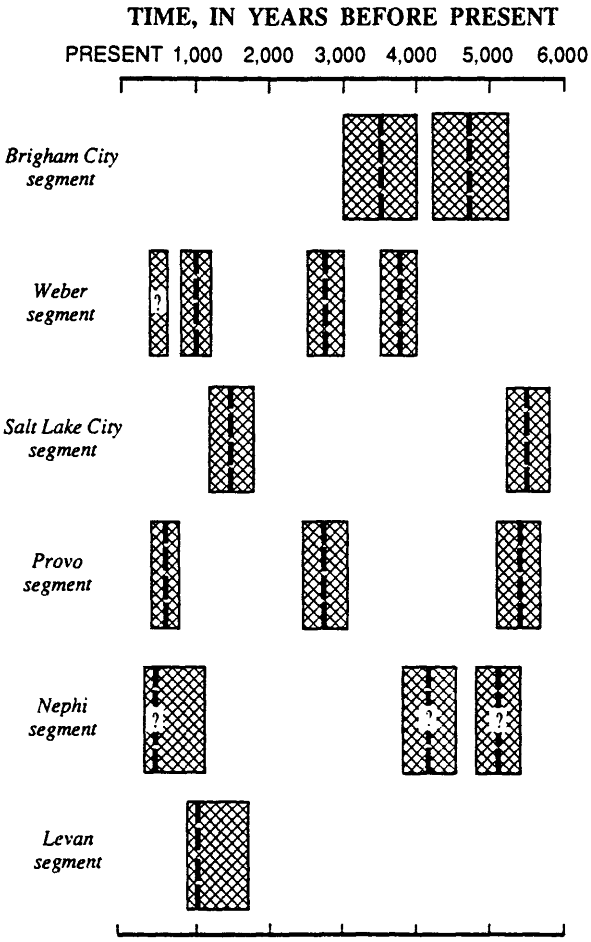

Fig. 4. Timing of movement on segments of the Wasatch fault zone during the past 6000 years. Heavy-dashed line indicates our best estimates for time of faulting; cross-hachure pattern indicates likely time limits as determined from radiocarbon and thermoluminescence age estimates. (See Machette *et al.* in press, for catalog and discussion of dates used in determining times of fault movement.)

Holocene faulting patterns

Several interesting patterns evolve from the chronology depicted in Fig. 4. One pattern is the apparent random distribution of faulting events during the 1500–6000-year time interval. This apparent random pattern could be interpreted as one of northward-sweeping waves of fault activity. There could be three waves of activity, the first affecting the northern half of the WFZ from about 5500 to 3000 years ago. The second wave sweeps from south to north cross the WFZ from the Nephi segments (4200 years ago) to the Weber segment (1000 years ago), and the third wave (the most recent cluster of activity; 400–1000 years ago) extends from the Nephi through Weber segments in a more random fashion. Noticeably missing is movement on the Brigham City segment in the last two waves, which might be explained by proximity to the intersection of the V-shaped belt of tectonism and the WFZ. In addition, the Salt Lake City segment appears to have been inactive during the last (third) wave, perhaps owing to its characteristically large displacement per event (4.5 m at Dry Creek), and its inherently long recurrence interval (about 4000 years, Fig. 4) (Schwartz & Lund 1987). If this wave pattern is real, it might be associated with loading of adjacent segments owing to a small but regionally significant component of left-lateral slip, as indicated by the WNW orientation of least principal stresses in the eastern Basin and Range province (Zoback & Zoback 1980). The least principal stress orientation for the WFZ appears to be approximately E–W as indicated by slip indicators on bedrock fault scarps and from focal mechanisms of small earthquakes in the region (Zoback 1983). However, as previously noted in the discussion of the Borah Peak earthquake, some

Table 3. Historic surface-rupturing earthquakes in the vicinity of the Wasatch fault zone

Name (date of earthquake	Magnitude (number of events	Fault(s) (segment)	Rupture length (km) Total	>25 cm NVD	NVD (m) (and net slip, m) Max.	Aver.	Total slip (m) (and source)
Hansel Valley (12 March 34)	M6.6 (single?)	Hansel Valley	11.5	6e	0.5	0.2e	n.d.
Hebgen Lake (18 August 59)	M_S7.5	Red Canyon	23	21e	6.7	2.4	6.8 (SD)
	M_W7.3	Hebgen	14.5	11e	6.1	1.6	1.0 (SD)
	(double)	Both faults	35 ± 5	26–32		(2.1)	10 (net GD)
Borah Peak (28 October 83)	M_S7.3	Lost River	36 ± 3	26 ± 2e	2.7	0.8	1.5–2.2 (GD)
	M_W6.8–7.0 (single)	(Thousand Springs)			(2.9)	(1.0)	1.5 (SD)

Symbols: NVD, net vertical displacement; e, estimated value; n.d., not determined. Magnitudes: M, unspecified Richter; M_S, surface-wave; M_W, moment. Slip determinations: GD, geodetic data; SD, computed from seismic data. Data from following sources: Hansel Valley, Utah—Shenon 1936, Slemmons 1977, Doser 1989; Hebgen Lake, Montana—Witkind *et al.* 1962, Witkind 1964, U.S. Geological Survey 1964, Doser 1985, Doser & Smith 1985, Hall & Sablock 1985, Smith *et al.* in press; Borah Peak, Idaho—Doser & Smith 1985, Crone *et al.* 1987, Smith *et al.* in press.

faults that have a significant component of oblique-slip in the subsurface might not produce geomorphic or structural evidence of oblique-slip at the surface (Crone *et al.* 1987).

A second pattern is more striking. There is strong evidence for a recent period of temporal clustering of large earthquakes; i.e. a strong grouping of surface-rupturing earthquakes on the WFZ over a geologically short time interval. If the most recent event on the Salt Lake City segment occurred about 1500 years ago (Fig. 4), then between 400 and 1500 years ago movement occurred on five of the six segments of the WFZ that have been active in the Holocene. The recent clustering (six faulting events during an interval of 1100 years) indicates that one major surface-rupturing event occurred every 220 years. In contrast, we estimate that a surface-rupturing earthquake occurred once every 395 ± 60 years during the past 6000 years along the WFZ between Brigham City and Nephi (Table 2). In addition, we have no evidence for a major surface-rupturing earthquake on the WFZ during the past 400 years, which is the youngest time we allow for the Weber segment and our best estimate for the most recent event on the Nephi segment. These relations point strongly to a process of temporal clustering of large-magnitude earthquakes on the WFZ, but the process seems to be intermittent through time. This pattern of temporal clustering is similar to that of the central Nevada–eastern California Seismic Belt in the western part of the Basin and Range province, where 11 earthquakes of $M > 6.5$ have occurred since 1860 (dePolo *et al.* 1991). Although the time scale of the clustering is different—130 years vs 1100 years—we consider the central Nevada–eastern California Seismic Belt to be a historic analog for movement on the WFZ during the past 1500 years.

COMPARISONS WITH THE WASATCH FAULT ZONE

The empirical relations between historic earthquakes and normal faulting (Slemmons 1977, Bonilla *et al.* 1984) and recent studies of prehistoric faulting in the region are our best analogs for the expected nature of future surface rupturing during large-magnitude earthquakes on the WFZ.

Historic earthquakes and surface faulting

Large-magnitude ($M \geq 7$) earthquakes have occurred in two regions of the western interior of the United States in historic times: (1) an elongate NE-trending zone known as the central Nevada–eastern California Seismic Belt (Wallace & Whitney 1984, dePolo *et al.* 1991), which has been the locus of most of the historic surface faulting in the Basin and Range province; and (2) the northern part of the Intermountain Seismic Belt (ISB), which is characterized by abundant Holocene and late Pleistocene faulting, but only three historic earthquakes that were accompanied by surface rupturing. As originally defined by Smith & Sbar (1974), the ISB is the active seismic zone between the Colorado Plateau, middle Rocky Mountains and northern Rocky Mountains provinces to the east, and the extended terrain of the Basin and Range province to the west (Fig. 2). Arabasz *et al.* (1987) show the ISB as an arcuate belt of pronounced seismicity that extends from southern Nevada and northern Arizona to northwestern Montana and Idaho. The three $M > 6.5$ earthquakes associated with demonstrable surface rupturing in the northern ISB (Table 3) are: the 1959 M_S7.5 Hebgen Lake earthquake in southwestern Montana (Doser 1985), the 1983 M_S7.3 Borah Peak earthquake in central Idaho (Doser & Smith 1985, Smith *et al.* in press) and the 1934 M6.6 Hansel Valley earthquake in northern Utah (Doser 1989).

Faulting events in the central Nevada–eastern California Seismic Belt have included both normal dip-slip and oblique-slip (dePolo *et al.* 1991), whereas the faulting events in the ISB have been primarily dip-slip. Therefore, the following discussion focuses on ISB earthquakes and their relation to the WFZ.

The Hebgen Lake earthquake was a complex normal-faulting event that probably occurred along reactivated older (Laramide) faults (Witkind 1964). Doser's analysis of seismic data from the Hebgen Lake earthquake indicates a composite of two subevents 5 s apart on one or more S-dipping fault planes. Surface rupturing occurred on two faults during this earthquake: an average of 2.4 m of surface offset along 23 km of the Red Canyon fault, and an average of 1.6 m of surface offset along 14.5 km of the Hebgen fault (U.S. Geological Survey 1964, Witkind *et al.* 1964). Because the two traces overlap, the net rupture length for this earthquake is between about 29 and 38 km; Witkind (1964) reported a total rupture length of 35 km (Table 3). However, the occurrence of small-displacement ruptures in heavily forested terrain leads us to suspect that the rupture length for the Hebgen Lake earthquake was probably underestimated by at least several kilometers. In addition, about 1 m of sympathetic movement occurred along 3 km of the Madison fault, 15 km west of Hebgen Lake (U.S. Geological Survey 1964). A maximum of 6.1 m of surface offset occurred during the Hebgen Lake earthquake (Witkind *et al.* 1962), with the bulk of the movement expressed as subsidence in the adjacent Yellowstone basin. In a more recent study, Hall & Sablock (1985) reported an average surface displacement of 2.1 m for the whole fault zone. Savage & Hastie (1966) estimated a maximum of 10 m of slip at depth from modeling of the geodetic data. In summary, it appears that the Hebgen earthquake, the largest recorded in the ISB, produced an average of several meters of surface rupturing along the two faults having a total length of 35 ± 5 km, and as much as 10 m of net slip at depth. The estimated seismic moment (M_0) (Aki 1966) for the main 1959 faulting event is 1.0×10^{27} N m, which equates to an estimated moment magnitude (M_W) of 7.3 (cited in Arabasz *et al.* 1987) using the empirical relations developed by Hanks & Kanamori (1979).

In October of 1983, the ISB was struck by the M_S7.3 Borah Peak, Idaho, earthquake. The earthquake was associated with 36.4 ± 3.1 km of surface rupturing that was concentrated along the Thousand Springs segment of the Lost River fault zone (Crone *et al.* 1987). Additional, subsidiary ruptures extended north on a basinward splay into the Warm Springs Valley and along the adjacent range-bounding Warm Springs segment of the Lost River fault zone. Of the total surface-rupture length reported for the earthquake, only 26 km (72%) had a continuous offset of ≥25 cm (Crone *et al.* 1987, fig. 4). An average of 0.8 m of vertical offset (1.0 m net slip) occurred along the Thousand Springs segment (Table 3). Geodetic data suggests about 1.56 m of offset at the surface and as much as 2.2 m of net slip along the fault at depth (Stein & Barrientos 1985), whereas body-wave modeling indicates about 1.5 m of slip (Doser & Smith 1985). Measurement of slickenlines and grooved surfaces on the exposed fault plane indicated an average of 17% left-lateral slip, whereas the preferred focal plane solution suggested about 30% left-lateral slip (Crone *et al.* 1987). The estimated M_0 for the 1983 faulting event is 2.1×10^{26}–3.1×10^{26} N m, which equates to an estimated M_W6.8–7.0 (cited in Arabasz *et al.* 1987).

The smallest of the three historic surface-rupturing earthquakes in the ISB—the M6.6 1934 Hansen Valley, Utah, earthquake—produced about 11.5 km of surface faulting (Slemmons 1977). This faulting occurred mainly within the basin floor of the valley (Shenon 1936) and was not associated with a major range-bounding fault. The fault's recurrence interval and slip rates appear to be an order of magnitude less than that of the WFZ (McCalpin *et al.* 1987, Doser 1989). The maximum displacement was 52 cm (Slemmons 1977), but the average was probably ≤20 cm; it appears that this earthquake was only slightly above the threshold magnitude for normal surface faulting, which is probably M_L6.3 ± 0.2 (Arabasz *et al.* 1987) or M_S6.0–6.25 (Doser 1989) for this portion of the Basin and Range province.

Prehistoric surface faulting in the region

One way to estimate the magnitude of prehistoric earthquakes on the WFZ is to compare its paleoseismologic parameters (length and offset) with historic surface-rupturing faults in the region using the empirical relations derived by Slemmons (1977) or Bonilla *et al.* (1984). However, all but three of the historic surface ruptures in the interior of the western United States are associated with the central Nevada Seismic Belt, and the majority of these Nevada earthquakes have large components of oblique-slip and significant lengths of small vertical displacement (<25 cm) that have produced scarps which are easily obliterated. Because the relations between length and offset along surface ruptures and earthquake magnitude are different for normal dip-slip vs strike-slip faulting (see Slemmons 1977, Bonilla *et al.* 1984), a more reasonable comparison might be made with the historic surface faulting that occurred closer to the WFZ and within the ISB. In addition, comparisons between the WFZ's paleoseismic data and segmented latest Pleistocene and Holocene faults in the northern part of the ISB can provide valuable insight into the nature and style of segmentation of normal faults in the region.

As previously mentioned, the northern part of the ISB is largely coincident with a belt of young faults (<15 ka) that form a right-stepping en échelon pattern from the northern part of the WFZ northeastward through the Cache, Bear Lake and Star Valleys, and from Jackson Hole to the Yellowstone area (Fig. 1); from Yellowstone, it trends westward and includes the 1959 movement on the Hebgen and Red Canyon faults, and young movement on the Centennial fault. The ISB extends across southwestern Montana and central Idaho as a corridor of NNW-striking, young range-bounding fault zones that typically have their highest slip rates, most recent movement, and maximum throw along the medial part of the fault zones (see Crone & Haller 1991).

The regional studies of late Quaternary surface faulting along the belt of young tectonism suggest a common range of segment lengths for range-bounding faults (Table 4) believed to be associated with large-magnitude (M ⩾ 7) earthquakes. Figure 5 summarizes the lengths of segments found in two classes of faults: (1) those along

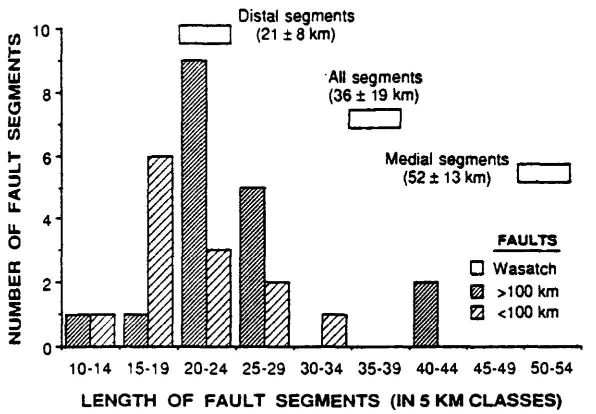

Fig. 5. Histogram of proposed segment lengths for late Quaternary fault zones that have been active in the past 15 ka in the northern Basin and Range province and northern part of the Intermountain Seismic Belt. Average length of segments: 24.6 ± 7.6 km on fault zones >100 km long; 19.8 ± 6.4 km on fault zones <100 km long. Lengths of various types of Wasatch fault zone segments are shown for comparison.

major range fronts that are typically >100 km long (the Beaverhead, Lemhi, Lost River and Wasatch fault zones) (Fig. 1) and (2) those along lesser range fronts that are typically <100 km long (the East Bear Lake, East Cache, Red Rock, Star Valley and Teton faults) (Fig. 1). The proposed segments on the longer faults range from 18 to 43 km in length and average about 25 ± 8 km, whereas the shorter faults have segments that range from 11 to 32 km in length and average about 21 ± 6 km (Table 4).

Some of the differences between the length of segments along historic and prehistoric faults probably can be explained by differences in slip-rate, slip-orientation and scale of mapping. For example, careful mapping of modern surface ruptures, such as those that formed during the 1983 Borah Peak earthquake (Crone *et al.* 1987, fig. 4), shows that as much as one-quarter of the length (9 km) may be of small (<25 cm) displacement (Table 3). Small ruptures might not be recognized along prehistoric faults solely on the basis of surficial geologic mapping or trench studies; thus, ancient rupture lengths may be underestimated in comparison with historic faults. If small-displacement ruptures overlapped an adjacent fault segment, even careful trenching may fail to detect a second faulting event on the adjacent segment (and vice versa).

Magnitude estimates for prehistoric WFZ earthquakes

Seismological and geological studies of three recent earthquakes suggest that rupture lengths ranging between 11 and 40 km have been associated with historic earthquakes of M_S6.6–7.5 (M_w6.8–7.3) and have primarily normal dip-slip displacement (Table 3). For comparison, studies of late Pleistocene and Holocene normal dip-slip faulting in the same region suggest that segment lengths average between about 20 and 25 km for prehistoric faulting along major uplifted ranges (Table 4). However, the WFZ, which is by far the longest normal fault zone in the western United States, is characterized by segments that are about twice as long (average length 52 ± 13 km) on the central active part and about the same (21 ± 8 km long) on the distal, less active portions (Table 1).

Holocene surface displacement on the medial segments of the WFZ typically has averaged between 2 and 3 m, with a maximum of about 4.5 m. Thus, the length (40–70 km) and surface-displacement (2–3 m) values for the medial segments of the WFZ are as large as those associated with historic faulting in the region. Both the Hebgen Lake and Borah Peak earthquakes occurred at depths of about 15 km (Smith *et al.* in press), which relate to a width of about 20 km on a fault plane that dips 50–60°. If you assume a similar geometry and depth for the WFZ, the seismic moment (M_0, Aki 1966) would have been 5.3×10^{26}–13.9×10^{26} dyne-cm for the prehistoric earthquakes on the WFZ. Using the relation of $M_W = (2/3) \log M_0 - 10.7$ (from Hanks & Kanamori 1979), the seismic moment converts to M_W7.1–7.4. The lesser value reflects 2 m of surface displacement on a 40-km-long segment (i.e. Brigham City or Nephi), whereas the greater value reflects 3 m of surface displacement on a 60–70-km-long segment (i.e. Weber or Provo). Using

Table 4. Length of proposed segments on late Quaternary fault zones, northern part of the Intermountain Seismic Belt

Fault zones >100 km long	Lengths (km, N to S) ($\bar{x}$ = 24.6 ± 7.6 km)*	Segmentation reference
Lost River, Idaho (141 km)	25, 18, 22, 22, 29, 25 ($\bar{x}$ = 23)	Crone & Haller (1991)
Lemhi, Idaho (150 km)	23, 23, 12, 43, 29, 20 ($\bar{x}$ = 25)	Crone & Haller (1991)
Beaverhead, Idaho (151 km)	20, 20, 23, 21, 42, 25 ($\bar{x}$ = 25)	Crone & Haller (1991)
Wasatch, Idaho and Utah (383 km)	17, 19, 30, 40, 61, 46, 69.5, 42.5, 30, 11 ($\bar{x}$ = 36)	This report, Machette *et al.* in press
Fault zones <100 km long	**Lengths (km, N to S) ($\bar{x}$ = 20.7 ± 5.6 km)**	**Segmentation reference**
Red Rock, Montana (27 km)	11, 16 ($\bar{x}$ = 14)	Stickney & Bartholomew 1987, Crone & Haller 1991
Teton, Wyoming (70 km)	24, 20, 20 ($\bar{x}$ = 21)	Susong *et al.* 1987, Byrd *et al.* 1988, Smith *et al.* 1990
Star Valley, Idaho (40 km)	24, 16 ($\bar{x}$ = 20)	Piety *et al.* 1986, Piety 1987
East Cache, Utah (55–62 km)	26, 15, 14–23 ($\bar{x}$ = 18–21)	McCalpin 1989
East Bear Lake, Utah & Idaho (≥78 km)	≥20, 26, 32 ($\bar{x}$ ≥ 26)	McCalpin 1990

*Wasatch fault zone not included in calculation of length.

similar parameters, Arabasz *et al.* (1987) calculated a maximum magnitude of M_S7.5–7.7 for the WFZ. Clearly, the WFZ poses a viable hazard to the urban population of Utah, both in terms of the recurrence and size of large-magnitude earthquakes.

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Appendix E

Document (Arabasz, 1991) presented to NEPEC by Allison outlining a consensus view of activities that would reduce losses due to earthquakes.

A GUIDE TO REDUCING LOSSES FROM FUTURE EARTHQUAKES IN UTAH "CONSENSUS DOCUMENT"

Walter J. Arabasz
Editor

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UTAH GEOLOGICAL AND MINERAL SURVEY
606 Black Hawk Way
Salt Lake City, Utah 84108-1280

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A Guide to Reducing Losses from Future Earthquakes in Utah

"Consensus Document"

Editorial Note: This document represents a consensus view of scientists, engineers, planners, emergency management officials, and others involved in a five-year program (1983-1988) as part of the National Earthquake Hazards Reduction Program focusing on earthquake hazards and risk in Utah. It was developed in connection with the "Fifth Annual Workshop on Earthquake Hazards and Risk Along the Wasatch Front, Utah" (January 31-February 2, 1989, Salt Lake City). —WJA

July 1989

(Revised June 1990)

FOREWORD

(Purpose and Contents)

The purpose of this document, as originally conceived, was to motivate and guide actions that will reduce losses from future moderate-to-large (magnitude 5.5 to 7.5) earthquakes in Utah, with primary emphasis on Utah's densely populated Wasatch Front region. In its present form, this document is viewed as an "intermediate-stage" product.

Public officials and decisionmakers in Utah need understandable and reliable information about Utah's earthquake threat. To meet their needs, it seems inescapable that one or more derivative documents—illustrated and simplified to meet the particular needs at hand—will have to be created. For example, Appendix B is a pamphlet entitled "Utah's Earthquake Threat" prepared in February 1990 for an earthquake-preparedness exposition at the Salt Palace (attended by more than 10,000 people). A book for the general public entitled "The Earthquake Threat—and Challenge—in Utah," currently being written by W. J. Arabasz and D. R. Mabey and sponsored by the U.S. Geological Survey, will be published in 1991 by the Utah Geological and Mineral Survey. The consensus view of scientists and engineers summarized in this document provides underpinnings for the book.

There are three basic parts to *this* document. Part One considers the question whether Utah is ready to take action to reduce its earthquake risk—and it is argued that seven key ingredients now exist for timely action in Utah. In Part Two, four basic strategies are outlined for communities in Utah to reduce earthquake losses. In Part Three, information is presented summarizing the nature and extent of the physical effects and losses that can be expected from earthquakes in Utah. This summary is based on up-to-date information and represents the consensus judgment of scientific and engineering experts involved in studies of Utah's earthquake problems. Technically-worded statements prepared by the scientists and engineers are presented in the Appendix A. A "layman's distillation" of those statements appearing in Part Three was written by S. J. Nava and W. J. Arabasz.

Walter J. Arabasz
Salt Lake City
June 1990

PART ONE

IS UTAH READY TO TAKE ACTION TO REDUCE ITS EARTHQUAKE RISK?

by
Genevieve Atwood[1] and Walter W. Hays[2]

YES—focused efforts during the last five years have achieved several successes including an adequate scientific and engineering base upon which to take action; a general willingness of public and private leaders to act responsibly relative to the earthquake risk; a general willingness of the public to accept actions to reduce the risk; and a willingness of a few key leaders—but not yet many elected officials—to provide leadership to bring about actions.

Key ingredients that now exist for future success in implementing earthquake hazard reduction include:

1. **A High Level of Concern**—Technically-trained public officials have an understanding of the earthquake hazards in Utah and realize that actions taken now can mitigate the hazard and reduce losses. The Wasatch Front news media is remarkably well-informed and has played a major role in enlightening the public to earthquake risks. Opinion polls show that the general public recognizes the potential for earthquake disasters and will support the adoption of a number of earthquake mitigation measures.

2. **Reliable Information**—Scientists, engineers, planners, and emergency response officials have amassed a substantial body of technical information about the Wasatch fault and other active faults in Utah—their location and geometry, the hazards associated with them, the recurrence of large earthquakes, and what actions will be effective in reducing the risk. New hazard maps and recent loss studies show the nature and extent of the earthquake threat along the Wasatch Front. This information clearly demonstrates the vulnerability of the region's economy to earthquake losses.

3. **User-Friendly Products**—A wide range of data, reports, maps, guidelines, and digitized information has been translated into plain wording to answer the basic questions asked by planners, emergency managers, and public officials—i.e., Where? How often? What effects? "Translated" hazard maps have been developed specifically for technical users and disseminated through the cooperative effort of federal, state, and local governments working together with the academic and private sectors. The county geologist program has been exceptionally effective in bridging the gap between information producers and information users in local government.

1. former Utah State Geologist, currently with Atwood & Mabey, Inc.
2. U. S. Geological Survey

4. **Professional and Institutional Support**—A core group of individuals believes the earthquake threat is real, and these individuals are trained and committed to devising effective and appropriate hazard reduction techniques for the Wasatch Front. This group includes social scientists, architects, planners, civil engineers, structural engineers, earth scientists, public decisionmakers, public-safety professionals, and business people. These individuals provide leadership within their own groups and exert influence beyond their organizations.

5. **Policy Champions**—Dedicated proponents of earthquake safety both within and outside Utah have promoted specific earthquake safety policies in Utah. Past experience has taught many lessons in how to succeed with decisionmakers, business people, and the public. Although Utah lacks sufficient public concern to force action and compel elected officials at all levels of government to make a crusade of the issue, decisionmakers do recognize earthquake hazard reduction as part of their responsibility for the public health, safety, and economic well-being of their communities.

6. **Information Exchange**—A network of information exchange links seismologists, structural engineers, and land-use planners. New findings in seismology, geology, and engineering can be readily transferred for incorporation into local hazard mitigation policies. Conversely, special needs in local policy can be readily addressed by experts drawing on an existing knowledge base. The network of information exchange enhances the credibility of mitigation policy even when implemented in a context of changing needs and expanding knowledge. New information can be incorporated into existing siting design, construction, retrofitting, and land-use practices by redefining map boundaries and refining existing concepts about the hazard without jeopardizing the fundamental credibility of the program.

7. **Window of Opportunity**—Will it take a major destructive earthquake before Utah takes significant actions to reduce its earthquake risk? Not now! Significant steps already have been taken (e.g., hospital construction standards, enactment of zoning ordinances), and other steps are ready to be taken. The damage caused by the 1982-86 wet cycle significantly increased (1) the level of awareness and (2) the commitment of public officials to make the state less vulnerable to geologic hazards.

The "window of opportunity" during which communities can accelerate the adoption of seismic safety measures is wide open in Utah. The most recent, comprehensive five-year effort, involving several hundred worker-years and more than 15 million dollars of federal, state, and local resources, was built upon the legacy of Utah's Seismic Safety Advisory Council and earlier regional seismological research. Now that most of the technical and societal information is in place, Utah is ready to take political and policy actions to reduce its earthquake risk.

PART TWO

BASIC STRATEGIES FOR LOSS REDUCTION

by
M. Lowe[1], G.E. Christenson[1], C.V. Nelson[2],
R.M. Robison[3], and J. Tingey[4]

To reduce its vulnerability to earthquakes, a community must adopt four basic strategies to keep expected losses within acceptable limits. These strategies necessarily involve an understanding of the earthquake threat, a knowledge of what actions will be effective in reducing risk, and an appreciation of the willingness and ability of the people involved to take action. The four basic strategies, which can be adopted and tailored to local needs, are: (1) improved development and construction practices; (2) public education concerning earthquake hazards and how to respond during a hazard event; (3) disaster-response plans; and (4) post-earthquake recovery plans.

Improvement of development and construction practices is primarily the responsibility of state, county, and municipal government agencies through adoption and enforcement of building codes and subdivision zoning, and retrofit ordinances. When faced with earthquake hazards, communities have five possible alternative actions: (1) ignore the hazard; (2) avoid the hazard; (3) modify the hazard (reduce the likelihood or severity of the hazard); (4) modify what is at risk (strengthen structure to withstand the hazard event); and (5) understand the hazard and accept the risk (usually involves disclosure of the hazard to potential owners and occupants) (Anderson, 1987).

Ignoring the hazard is not an acceptable action as it does not fulfill government's mandate to protect the health, safety, and welfare of its citizens and may lead to governmental liability for damages and/or loss of life accompanying earthquakes. In determining which of the other alternative actions is most appropriate, the risk, in terms of both economic and life loss, should be considered along with the cost of avoiding or mitigating the hazard and the type of facility which is being considered. Table 1 lists typical hazard-reduction techniques for some of the more widespread types of earthquake hazards. Which techniques are most appropriate for a particular development must generally be determined by a site-specific study.

1. Utah Geological and Mineral Survey
2. Salt Lake County Geologist
3. former Utah County Geologist; currently with Sergent, Hauskins, and Beckwith
4. Utah Division of Comprehensive Emergency Management

One of the more serious problems in promoting earthquake-hazard reduction is convincing the public that there is indeed a hazard. (Only a few of Utah's urban areas have been damaged historically by close earthquakes of Richter magnitude 5.5 or larger [Richfield, magnitude 6½, 1901; Elsinore, two shocks of magnitude 6, 1921; Logan, magnitude 5.7, 1962].) In order to show the need for taking steps to reduce earthquake hazards, technical information must be translated so that it may be understood by the layman. This translated information must identify the likelihood of occurrence, location, severity in terms of what will happen when the event occurs, and what steps may be taken to reduce the risk. This consensus document is one attempt to provide translated information about earthquake hazards to the layman.

The purpose of disaster-response plans is to identify: (1) the types of decisions that are likely to be needed when the expected earthquake event occurs, (2) who will make the decisions, and (3) how the decisions will be transmitted to the public and emergency-response personnel so that they may be implemented. Disaster-response exercises are conducted so that implementation of disaster-response plans will occur in the fastest, most efficient manner possible.

Recovery plans are designed to anticipate and meet the time-varying needs of the community as the post-earthquake recovery period unfolds over a period of 5 to 10 years. These plans will help ensure that the community quickly returns to cultural and economic viability following an earthquake.

Basic products are now, or soon will be, available to develop and carry out these strategies for earthquake-loss reduction in Utah. They include: (1) maps showing susceptibility to earthquake hazards such as ground shaking, surface rupture, slope failure, and liquefaction, and depicting either explicitly or implicitly the affected area, severity of impact, frequency of occurrence, impact time, duration, and the potential for triggering secondary effects; (2) loss studies identifying the distribution and nature of the damage and losses expected in the realistic scenario of one or more earthquakes; and (3) risk-reduction studies based on experience in Utah communities and elsewhere describing which risk-reduction actions are likely to be most effective.

TABLE 1

PRINCIPAL EARTHQUAKE HAZARDS, EXPECTED EFFECTS, AND COMMONLY-APPLIED TECHNIQUES TO REDUCE HAZARDS

Hazard	Expected Effects	Commonly Used Hazard-Reduction Techniques. Other Mitigation Techniques May Be Used Which Are Not Listed Here.
Surface-Fault Rupture	Rupture of ground with relative displacement of surface up to 20 feet along main trace of fault. Tilting and ground displacements may occur in a zone of deformation up to several hundred feet wide, chiefly on the downthrown side of the main fault trace.	Avoid active fault traces by setting structure back a safe distance from fault.
Ground Shaking	Vertical and horizontal movement of the ground as seismic waves pass. Damage or collapse of man-made structures can result, depending on the amplitudes, frequencies, and duration of ground motions. Horizontal motions generally cause greatest damage. Damaging ground motions can occur as far as 60 miles from the earthquake source, depending on source, path, and site conditions.	Design and build new structures to meet or exceed the seismic provisions in the current Uniform Building Code. Replace or retrofit older structures (especially unreinforced masonry buildings) to strengthen them so they meet current UBC requirements. Tie down water heater and secure heavy objects inside buildings.
Tectonic Subsidence	Regional tilting of valley floor toward fault causing flooding near lakes and in areas of shallow ground water. May cause loss of head in gravity-flow structures (for example, sewer systems).	Increase tolerance for tilting in gravity-flow structures; design structures for releveling. Buffer zones or dikes around lakes or impounded water to limit flood hazard; prohibit basements in shallow ground-water areas.
Liquefaction	Water-saturated sandy soils may liquefy (become like quicksand) causing differential settlement, ground cracking, subsidence, lateral downslope movement of upper soil layers on gentle slopes, and flow failures (landslides) on steep slopes.	Improve soil-foundation conditions by removing susceptible soils, densification of soils through vibration or compaction, grouting, dewatering with drains or wells, and loading or buttressing to increase confining presures. Structural solutions include use of end-bearing piles, caissons, or fully compensated mat foundations.
Earthquake-Induced Rock Fall	Downslope movement of bedrock fragments and boulders causing damage due to impact.	Avoidance. Remove or stabilize potential rock-fall sources by bolting, cable lashing, burying, or grouting. Protect structures with deflection beams, slope benches, or catch fences.
Earthquake-Induced Landslides	Downslope movement of earth material causes damage to structures below the landslide due to impact and/or burial. Differential displacement of scarps and movement in both vertical and horizontal directions causes loss of foundation support for structures within and adjacent to the central mass of the landslide.	Avoidance. Remove landslide-prone material. Stabilize slopes by dewatering, retaining structures at toe, piles driven through landslide into stable material, weighting, or buttressing slopes. Bridging.
Earthquake-Induced Seiches	Earthquake-generated water waves causing inundation around shores of lakes and reservoirs. Loss of life due to drowning. Damage due to flooding, erosion, and pressures exerted by waves.	Avoidance. Flood-proofing and strengthening to withstand wave surge. Diking. Elevate buildings.

PART THREE

THE EARTHQUAKE THREAT IN UTAH

A Consensus on the Expected Physical Effects and Potential Losses Associated With Future Earthquakes

As early as 1883, the eminent geologist G.K. Gilbert recognized and warned of the serious earthquake threat posed by the Wasatch fault and other active faults in Utah despite the absence up to that time of any large earthquakes in the region since settlement by Mormon pioneers in 1847. In modern times, seismologists, geologists, and engineers, have amassed a large body of technical information and have reached fundamental agreement about Utah's earthquake dangers. That consensus was arrived at as part of a special five-year focus (1983-1988) on the Wasatch Front region under the National Earthquake Hazards Reduction Program. The technical "consensus" statements of the scientists and engineers are presented verbatim in the Appendix of this document. A "layman's distillation" of the technical information is summarized here. Numbered references appearing in the margins are keyed to the Appendix. It should be noted that the following summary and appended statements reflect a general agreement on what to expect, even though some scientific and technical issues may not be fully resolved.

Our understanding of earthquake danger in Utah is based on earthquakes experienced in Utah, earthquakes that have occurred elsewhere in the western United States, our knowledge of the geology of Utah, and research on earthquake mechanisms and effects. This understanding has led most, if not all, scientists who have studied the problem to conclude that the Wasatch Front area, where 90 percent of Utah's population resides, is an active seismic zone with earthquake dangers that demand the attention of officials and the general public. Although a destructive earthquake could occur anywhere in Utah, the primary focus of this discussion—because of the large population at risk—will be a large surface-faulting earthquake on the Wasatch fault and physical effects that are expected in the eleven counties within or adjacent to the Wasatch Front: Salt Lake, Davis, Juab, Weber, Wasatch, Summit, Morgan, Cache, Utah, Tooele, and Box Elder. However, many of the general statements presented in this document are also applicable for earthquakes occurring elsewhere in Utah than on the Wasatch fault. Utah's earthquake problems emphatically are not restricted to the Wasatch fault.

Reference (Keyed to Appendix)	General Statements

The state of Utah is transected by the Intermountain seismic belt, a coherent northerly-trending belt of **earthquake** activity extending at least 1,500 kilometers (900 miles) from southern Nevada and northern Arizona to northwestern Montana. The Intermountain seismic belt is characterized by shallow scattered earthquakes less than 25 kilometers (15 miles) deep, geologically active **normal faults**, and high **seismic risk** associated with episodic **surface-faulting** earthquakes of about **magnitude** 6.5 to 7.5.

[1.2] **Seismic hazards** in the Wasatch Front arise from the potential for two different types of earthquake occurrence: (1) Moderate-sized earthquakes that are not constrained in location to mapped faults and that may occur anywhere throughout the region, and (2) infrequent large surface-faulting earthquakes on identifiable faults having evidence of geologically recent movement.

[1.2.2] Moderate but potentially damaging non-surface-faulting earthquakes (magnitudes 5.5 to 6.5) may occur anywhere within the Wasatch Front region. These earthquakes may occur on either known or unknown faults. Unknown faults include buried faults which cannot be seen at the surface. Based upon instrumentally-
[1.3.2] recorded earthquakes since 1962, potentially damaging earthquakes of magnitude 5.5 and larger are expected to occur in the Wasatch Front region about once every 14 to 40 years. Eight earthquakes with measured or estimated magnitudes of 5.5 or greater occurred in this region from 1850 through 1988, the most recent being the 1975 Pocatello Valley earthquake near the Utah-Idaho border.

Moderate-sized earthquakes have the potential to produce substantial damage in the Wasatch Front urban corridor. A "direct-hit" to one of the Wasatch Front's

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earthquake: A sudden trembling in the earth caused by slippage on a fault (fracture) accompanied by the abrupt release of slowly accumulated strain energy.
fault: A fracture or fracture zone along which there has been displacement of the sides relative to one another along the surface of the fracture.
normal fault: A fault whose movement is primarily in a vertical direction. The Wasatch fault is an example of a normal fault with the mountain block rising relative to the valley floor.
seismic risk: The social or economic consequences of future possible earthquakes. Risk may be expressed as the probability that adverse effects will equal or exceed specified values in an area during a specified interval of time.
surface faulting (surface rupture): Displacement of the ground surface by a fault movement. Surface faulting is expected to occur in Utah only in earthquakes of magnitude 6.3 or larger. The majority of small to moderated-sized earthquakes in Utah occur on faults whose rupture does not reach the surface.
magnitude: A number that characterizes the size of an earthquake from measurable motions recorded by a seismograph, corrected for the distance to the source of the earthquake.
seismic hazard: Any physical phenomenon (e.g., ground shaking, ground failure, surface-faulting) associated with an earthquake that may produce adverse effects on human activities.

major cities could result in more than $2.3 billion for a magnitude 6.5 earthquake—and more than $830 million for a magnitude 5.5 earthquake. A magnitude 6.5
[2.1.4] earthquake is expected to produce ground motions on soils within 10 kilometers (6 miles) of the fault that range from at least 0.25 to 0.5g. [The force of gravity is 1.0g.] At 20 kilometers (12 miles) from the fault, corresponding ground motion estimates on soil are 0.15 to 0.25g. The greater ground shaking will tend to occur at sites underlain by unconsolidated alluvium and lake deposits, and lesser ground
[2.3.3] shaking will tend to occur at sites underlain by rock. Ground motion levels at soft sediment sites are expected to be 6 to 10 times greater than at rock sites.

[3.1.3] Earthquakes of magnitudes less than 6.5 could cause rock falls, rock slides, and other slope instabilities within a few miles of the earthquake source. Such
[3.1.1] earthquakes also could trigger **liquefaction** locally.

Earthquakes of about magnitude 6.3 and greater occurring along the Wasatch
[1.5.1] Front are expected to produce surface-faulting. Since 1850, there have been three historical earthquakes in the Intermountain seismic belt that were associated with documented surface-faulting:

Year	Magnitude	Location	Maximum Surface Displacement
1934	6.6	Hansel Valley, Utah	0.5 meters (1.6 feet)
1959	7.5	Hebgen Lake, Mont.	5.5 ± 0.3 meters (18.0 ± 1.0 feet)
1983	7.3	Borah Peak, Idaho	2.7 meters (8.9 feet)

[1.4.1] Future large earthquakes on the Wasatch fault and other major faults in Utah are expected to have characteristics similar to those of large normal faulting earthquakes occurring in nearby states. These include: the 1959 Hebgen Lake, Montana, earthquake; the 1983 Borah Peak, Idaho, earthquake; and earthquakes up to magnitude 7.7 that have occurred in this century in Nevada.

The greatest threat for large surface-faulting earthquakes in Utah is posed by the Wasatch fault zone—despite the fact that it has not generated any earthquakes larger
[1.2.1] than magnitude 5 in historical time. There are many other known active faults in Utah that show evidence of prehistoric surface-faulting and that may produce large-surface-faulting earthquakes in the future. In general, the intervals between larger earthquakes on these faults tends to be considerably longer than for repeat ruptures along the most active parts of the Wasatch fault.

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liquefaction: The process by which water-saturated, unconsolidated sediments subjected to shaking in an earthquake temporarily lose strength and behave like a fluid. The lower areas of many of western Utah's valleys are susceptible to liquefaction.

The Wasatch fault zone follows the base of the western edge of the Wasatch
Range, from Malad City, Idaho, southward to Fayette, Utah, for a distance of
[1.4.4] 380 kilometers (240 miles). The Wasatch fault is made up of as many as 12
independent **segments**. Each segment is expected to rupture independently, although
rupturing on one segment may be followed closely in time by rupture on another
segment. This pattern of more-frequent-than-average rupturing is termed temporal
[1.1.1] clustering. The central 6 to 8 segments of the Wasatch fault zone (from Brigham
City to Levan)—based on trenching and dating studies—have repeatedly produced
magnitude 7.0 to 7.5 earthquakes during the past 6,000 years. The timing pattern of
[1.1.2] such large surface-faulting earthquakes on the Wasatch fault during the past 6,000
years is complicated. For the segments between Brigham City and Nephi, the
[1.3.1, composite recurrence interval—the average time between two faulting events
1.1.2] anywhere on this central part of the fault zone—ranges from a maximum of 415
years to a minimum of 340 years.

Based upon studies of **fault scarps** and the ages and timing of fault offsets, the
[2.1.1] probability of a large surface-faulting earthquake on the Wasatch fault in the next 50
years has been estimated to be between 4 and 20 percent. However, because of the
variability of the data used in this analysis and the possibility of multiple interpreta-
tions, it should be noted that a higher probability of occurrence in 50 years cannot
be ruled out.

Surface-faulting accompanying a magnitude 7.0 to 7.5 earthquake on the
Wasatch fault zone can be envisioned as follows. The fault rupture would likely
[1.4.2] extend beneath the valleys adjacent to the fault, probably to depths of 10 to 20
kilometers (6 to 12 miles). The length of the surface rupture would range from 20
to 70 kilometers (12 to 44 miles), depending upon the fault segment involved. A
[1.5.2, complex zone of faulting could be formed up to 500 meters (1640 feet) wide. Fault
1.4.2] scarp heights could be as much as 5 to 6 meters (16.5 to 19.5 feet).

The hazard from surface-faulting will be localized along a single segment of the
[1.5.4] fault—as opposed to the associated ground shaking and ground failure, which will
affect a much larger area and may be most intense away from the fault.

[2.1.3] The severity of ground shaking expected within urban areas adjacent to the rup-
tured fault segment (within 10 kilometers) roughly corresponds to a Modified

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segment: Portion of a fault that ruptures as a unit during an earthquake.
fault scarp: The cliff or steep slope formed by a fault that breaks the earth's surface.

Mercalli Intensity (MMI) of VIII on firm sediments and Modified Mercalli X or greater on soft sediments (0.4 to 0.6g). Modified Mercalli Intensity VIII is
[2.1.2, characterized by: partial collapse of weak masonry walls; fall of chimneys,
2.1.3] factory stacks, towers, elevated tanks; frame houses moved on foundations if not bolted down. Modified Mercalli Intensity X is characterized by: most masonry and frame structures destroyed with their foundations; some well-built wooden structures destroyed; serious damage to dams, diking and embankments; rails bent; and large landslides.

[3.1.1] Large earthquakes of magnitude 7.0 and greater are likely to trigger liquefaction and destructive ground failures in the sediments that lie beneath many areas in the
[3.1.2] lower parts of valleys along the Wasatch Front. The most common consequence of liquefaction along the Wasatch Front is expected to be **lateral spreading**. Lateral spreads would cause ground displacements of up to several feet, along with fracture of buildings, roads, and other surface works located on the unstable ground.
[3.1.3] Major damage from rock falls, rock slides, and other slope instability should be expected on steep slopes (such as within canyons and along mountain fronts) over a wide area.

In a magnitude 7.5 earthquake on a central part of the Wasatch fault, Utah
[2.4.1] should expect damage to buildings to exceed $4.5 billion in Davis, Salt Lake, Utah and Weber counties. This may represent only 20 percent of the total economic loss. Unreinforced masonry buildings (for example, brick homes built before 1960) are
[2.4.2] particularly vulnerable to ground shaking and are expected to account for 75 percent of the building losses. The Wasatch Front area has a sizable inventory of other structures not built with earthquake-resistant design that will be seriously damaged.

Surface-faulting, and other ground failures due to ground shaking during a large earthquake, will cause major disruption of lifelines (utilities, water, sewer), transportation systems (highways, bridges, airports, railways), and communication systems.
[2.4.4] As a result of the geographical concentration of state-owned buildings—and their limited seismic resistance—losses from a large Wasatch fault earthquake could easily reach 30 or 40 percent of replacement value. (Schools, hospitals, and fire stations were not studied.)

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lateral spreading: Landslides that form on gentle slopes as the result of liquefaction of a near-surface layer from ground shaking in an earthquake.

[2.4.3] A 1976 study by the U.S. Geological Survey for a worst-case earthquake on the central Wasatch fault estimated 2,300 fatalities (assuming no dam failures), 9,000 injured, and 30,000 homeless. The number could be as high as 14,000 if deaths from dam failure are included in the casualty total. The experience of the 1988 Armenian earthquake—and more up-to-date engineering judgment about the collapse potential of many structures in the Wasatch Front area—suggests the 1976 fatality estimate is low.

[2.4.6] There may be losses relating to disturbance of the Great Salt Lake and Utah Lake from a major earthquake. The magnitude of the losses would be dependent upon such factors as lake elevation and the amount of downward tilting of a valley floor toward the fault scarp. Rapid inundation of developed areas adjacent to lakes could result in large losses of life and property. **Seiches** may cause the Great Salt Lake and Utah Lake to oscillate for many hours, temporarily raising and lowering the water level. Additional losses could be expected from flooding due to possible failures of dams or other water impoundment structures and from fires.

seiche: Oscillations (standing waves) of the surface of a closed body of water when the surface is disturbed by wind or an earthquake.

APPENDIX A

Editorial Note: The summary statements that follow were written by small working groups formed during a planning meeting in Salt Lake City on November 9, 1988. A draft of the statements was distributed—and revised in response to group discussion—at the "Fifth Annual Workshop on Earthquake Hazards and Risk Along the Wasatch Front, Utah" (January 31-February 2, 1989, Salt Lake City). Sections 2.1 through 2.3 here, prepared by A. M. Rogers and others, consist of revised text written in September 1989. —WJA

1.0 HAZARDS ASSOCIATED WITH EARTHQUAKES AND SURFACE-RUPTURING FAULTS

1.1 The Wasatch Fault Zone (M.N. Machette, W.R. Lund, D.P. Schwartz, R.L. Bruhn)

1.1.1 Trenching and dating studies indicate that the southern 220 km (Brigham City to Levan) of the 343-km-long Wasatch fault zone is made up of 6 to 8 independent fault-rupture segments that have repeatedly produced M 7-7.5 earthquakes during the past 6,000 years (Machette and others, 1987). Siting and design criteria should be based on the expectation that the next large earthquake on the Wasatch fault will occur on one of these segments. (The next large earthquake in Utah, however, may not necessarily occur on the Wasatch fault; see 1.2.1.)

1.1.2 The pattern of timing of large surface-faulting earthquakes (M 7-7.5) on the Wasatch fault during the past 6,000 years is complicated. For the segments between Brigham City and Nephi, the *composite* recurrence interval—the average time between two faulting events anywhere on this central part of the fault zone—ranges from a maximum of 415 years to a minimum of 340 years (Machette and others, 1989).

1.1.3 Of the 6 to 8 active fault-rupture segments, the elapsed time since the last large earthquake has been longest on the Brigham City and Salt Lake City segments (3,600 and 1,500 years, respectively; Machette and others, 1989). However, the recurrence intervals for faulting on any one segment are usually quite variable (Machette and Scott, 1988, fig. 6).

1.1.4 Each segment of the Wasatch fault zone is expected to exhibit independent movement, although rupturing on one segment may be followed closely in time by rupture on another segment (Schwartz and Coppersmith, 1984; Machette and others, 1987). This pattern of more-frequent-than-average rupturing is termed temporal clustering. During an earthquake, most of the rupturing will be concentrated on the causative segment.

1.2 Sources of Seismic Hazard (J.C. Pechmann, M.N. Machette, W.J. Arabasz, K.M. Shedlock)

Seismic hazards in the Wasatch Front region arise from two different classes of earthquakes (Arabasz and others, 1987):

1.2.1 Large (M 6.3±0.2 to 7.5±0.2) earthquakes, accompanied by surface rupture, will occur in the future on the Wasatch fault as well as on a number of other known active faults in the region showing evidence of prehistoric surface faulting.

1.2.2 Moderate but potentially damaging earthquakes without surface rupture (M 5.5 to 6.5) may occur anywhere within the Wasatch Front region on either known or unknown faults. Unknown faults include buried faults which cannot be seen at the surface.

1.3 Frequency of Earthquake Occurrence (J.C. Pechmann, M.N. Machette, W.J. Arabasz, K.M. Shedlock)

1.3.1 Large surface-faulting earthquakes occur somewhere along the Wasatch fault on the average of once every 340 to 415 years (Machette and others, 1989). Large earthquakes are known to occur less frequently on other faults in the region for which information on earthquake recurrence is available (e.g., Youngs and others, 1987).

1.3.2 Analysis of the instrumental earthquake catalog from July 1962 through 1985 indicates the likelihood that potentially damaging earthquakes of M 5.5 or greater will occur on the average of once every 14 to 40 years in the Wasatch Front region. Eight earthquakes with measured or estimated magnitudes of 5.5 or greater occurred in this region from 1850 through 1988, the most recent being the 1975 M 6.0 Pocatello Valley earthquake near the Utah-Idaho border (Arabasz and others, 1987).

1.4 Characteristics of Future Large Earthquakes (J.C. Pechmann, M.N. Machette, W.J. Arabasz, K.M. Shedlock)

1.4.1 Future large earthquakes on the Wasatch fault and other major faults in Utah are expected to have characteristics similar to those of large normal faulting earthquakes that have occurred in nearby states. These earthquakes include: the 1959 M 7.5 Hebgen Lake, Montana, earthquake; the 1983 M 7.3 Borah Peak, Idaho, earthquake; and earthquakes of up to M 7.7 that have occurred this century in Nevada.

1.4.2 Future large Wasatch Front earthquakes could break sections of fault up to 70 km long and produce maximum vertical displacements at the surface of up to about 6 m. (Schwartz and Coppersmith, 1984; Machette and others, 1987; Arabasz and others, 1987). The fault ruptures will extend beneath the valleys adjacent to the faults, probably to depths of 10 to 20 km (Smith and Richins, 1984).

1.5 Surface Faulting (W.R. Lund, M.N. Machette, D.P. Schwartz)

1.5.1 Earthquakes having magnitudes of 6¼ (±¼) and greater along the Wastach Front are expected to produce surface faulting.

1.5.2 Surface faulting accompanying a magnitude 7.0-7.5 earthquake on the Wasatch fault zone will likely be characterized by:

(a) Rupture patterns that are expressed as a single fault trace or as several sub-parallel or branching traces that form a complex zone of faulting up to 500 m (1,650 ft) wide.

(b) Length of surface rupture 20-70 km (12.5-44 mi)

(c) Net tectonic displacement 2-5 m (6.5-16.5 ft)

(d) Scarp heights that can be as much as 5-6 m (16.5-19.5 ft) high with associated antithetic faulting, graben formation, and backtilting. The zone of intense ground deformation along *individual* fault traces can be as much as 50 m (165 ft) wide.

1.5.3 Surface faulting will destroy or severely damage lifelines (roads, utilities, pipelines, communication lines) that cross the fault as well as any structures built in the fault zone.

1.5.4 The hazard from surface faulting will be localized along a single segment of the fault—as opposed to associated ground shaking and ground failure, which will affect a much larger area and may be most intense away from the fault.

2.0 HAZARDS ASSOCIATED WITH GROUND SHAKING

2.1 Ground Motions for the Maximum Earthquake on the Wasatch Fault (A.M. Rogers, S.T. Algermissen, K.W. Campbell, D.M. Perkins, J.C. Pechmann, M.S. Power, J.C. Tinsley, T.L. Youd)

2.1.1 Rupture of one of the longer segments of the Wasatch fault could produce an earthquake as large as M = 7.5. Based on the studies of fault scarps and the ages and timing of fault offsets (Machette and others, 1989), the probability of such an event somewhere on the Wasatch fault in the next 50 years has been estimated to be between 4 and 20 percent (Perkins, personal comm.; Youngs and others, 1987); however, because of the variability of the data used in this analysis and the possibility of multiple interpretations, one should note that a higher probability of occurrence in 50 years cannot be ruled out at present.

2.1.2 For a M = 7.5 earthquake, urban areas adjacent to the ruptured segment (within 10 km) are expected to experience peak horizontal accelerations on soil sediments ranging from at least 0.4 to 0.6 g and peak horizontal velocities ranging from at least 50 to 100 cm/s. This zone includes most of the incorporated region of Salt Lake City, for example. At 20 km from the rupture, corresponding ground motion estimates on soil sediments are 0.2 to 0.3 g for peak acceleration and 25 to 50 cm/s for peak velocity (Campbell, 1987; Youngs and others, 1987; M.S. Power, written comm.)[1]. The motions on soft soil

1. It should be noted that the ground motion values quoted are based on data recorded primarily in California and are representative of ground motions from strike-slip faults. The Wasatch fault is a normal fault; theoretical studies suggest that a normal fault that dips underneath Salt Lake Valley would act to focus more energy in the urban area than might be expected from a strike-slip fault (Benz and Smith, 1988). This focusing is due both to the fact that the fault dips underneath the urban area and to the nature of rupture propagation on the fault.

sediments[2] will tend to be larger than on firm sediments, especially in terms of peak velocities[3].

2.1.3 The estimated values of ground motion for the hypothesized M = 7.5 earthquake roughly correspond to a Modified Mercalli Intensity (MMI) of VIII on firm sediments and X or greater on soft sediments. Modified Mercalli Intensity VIII is characterized by: partial collapse of weak masonry; damage to ordinary masonry; some damage to reinforced masonry; fall of some masonry walls; fall of chimneys, factory stacks, towers, elevated tanks; frame houses moved on foundations if not bolted down. Modified Mercalli Intensity X is characterized by: most masonry and frame structures destroyed with their foundations; some well-built wooden structures destroyed; serious damage to dams, dikes, and embankments; rails bent, and large landslides.

2.1.4 Smaller but more frequent earthquakes are expected to occur in the region that could also produce substantial damage in the Wasatch Urban Corridor (see the section on losses, this report). For example, M = 6.5 events are expected to produce ground motions on soil within 10 km of the fault that range from at least 0.25 to 0.5 g (25 to 75 cm/s). At 20 km from the rupture, corresponding ground motion estimates on soil are 0.15 to 0.25 g (15 to 35 cm/s) (Campbell, 1987; and Youngs and others, 1987; M.S. Power, written comm.). Again, the motions on soft soil sites will tend to be at the higher end of these ranges, especially for peak velocity.

2.2 **Probabilistic Ground Motion Hazard** (A.M. Rogers, S.T. Algermissen, K.W. Campbell, D.M. Perkins, J.C. Pechmann, M.S. Power, J.C. Tinsley, T.L. Youd)

2.2.1 In any 50-year time period, there is a 10 percent probability that the levels of peak horizontal ground acceleration and velocity at sites underlain by firm sediments will exceed the range 0.20 to 0.35 g and 20 to 50 cm/s, respectively, along the Wasatch Front (Algermissen and others, in preparation; Youngs and others, 1987; M.S. Power, written comm.). These values are most likely to occur within a 10-km zone located to the west of the surface trace of the Wasatch fault. These values are based on the contemporary Wasatch Front region seismic record and evidence of large earthquakes in the recent geologic past. The estimates incorporate the effects of earthquakes on the Wasatch fault, as well as more distant earthquakec. The ground motions cited correspond

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2. The term "soft sediments" is used collectively to refer to sediments of low near-surface shear velocity, high near-surface shear velocity gradients, and high shear velocity contrast at the base of the sediments, which tend to occur in those parts of Salt Lake Valley underlain by deep sediments.

3. One should also note that studies by Benz and Smith (1988) indicate that at least half the spectral amplification observed in Salt Lake Valley, at periods greater than about 0.7 seconds, can be attributed to the velocity contrast between basin sediments and crystalline basement as opposed to amplification associated with near-surface soft sediments.

roughly to Modified Mercalli Intensity VIII to IX. It is likely that at this same probability level some sections of the urban areas near the epicenter would experience ground motions larger or smaller than these values, reflected by intensities one to two units above or below VIII. Higher ground motions and damage will tend to occur at sites underlain by unconsolidated alluvium and lake deposits, and lower damage levels will tend to occur at sites underlain by rock.

2.2.2 In any 10-year time period, there is a 10 percent probability that the levels of peak horizontal ground acceleration and velocity at sites underlain by firm sediments will exceed 0.06 to 0.08 g and 5 to 9 cm/s, respectively, along the Wasatch Front (Algermissen and others, in preparation; Youngs and others, 1987; M.S. Power, written comm.). These values are likely to occur anywhere within the Ogden-Salt Lake-Provo corridor. These ground motions correspond roughly to Modified Mercalli Intensity IV to VI.

2.2.3 In any 250-year time period, there is a 10 percent probability that the levels of peak horizontal ground acceleration and velocity at sites underlain by firm sediments will exceed 0.5 to 0.7 g and 55 to 110 cm/s, respectively, along the Wasatch Front (Algermissen and others, in preparation; Youngs and others, 1987; M.S. Power, written comm.). These values are most likely to occur within a 10-km zone located to the west of the surface trace of the Wasatch fault and correspond roughly to Modified Mercalli Intensity IX to X or greater.

2.2.4 Neither the deterministic ground-motion values based on maximum magnitude, the probabilistic ground-motion values, nor the intensities cited above are necessarily intended to be the design motions for this region. The choice of design ground motions should be based on the level of risk deemed appropriate for a given level of design motion. That is, a level of risk should be chosen that is acceptable to the engineering community and public officials for various classes of structures. Nevertheless, at 50-year exposure time, 10 percent probability of exceedance description of ground motion is consistent with that used by the Applied Technology Council (1978) for design ground-motion maps included in their proposed seismic regulations for buildings. The same specifications for ground motion are used in the National Earthquake Hazards Reduction Program "NEHRP Recommended Provisions for the Development of Seismic Regulations for Buildings" (Building Seismic Safety Commissions, 1985) and are the basis for the new 1988 Uniform Building Code (UBC). Thus, the probabilistic ground-motion values quoted above can be compared directly with ground-motion maps used nationally for the development of seismic provisions of building codes. The 10-year exposure period ground motions have not been used as design motions in the past, but are cited here to convey the short-term hazard, which is mostly due to intermediate-sized earthquakes. The 250-year exposure period ground motions have been used in the past as design values for critical facilities, such as hospitals, power plants, etc. For this exposure, the probabilistic ground motions convey the hazard due to the occurrence of large earthquakes, which are also more likely to occur over a 250-year exposure period compared to shorter intervals.

2.3 The Effect of Site Conditions on the Ground Motions of Distant Earthquakes (A.M. Rogers, S.T. Algermissen, K.W. Campbell, D.M. Perkins, J.C. Pechmann, M.S. Power, J.C. Tinsley, T.L. Youd)

2.3.1 Based on recordings of distant nuclear explosions in Nevada, it is known that sediment properties in Salt Lake Valley can produce substantial geographical variation in the level of ground motions (Hays, 1987; King and others, 1987). Theoretical studies of ground motion in Salt Lake Valley qualitatively support this observation (Benz and Smith, 1988; Schuster and others, 1990). The data collected by Hays and King, and others suggest that mean spectral estimates of low-amplitude ground-motion values are increased by factors of 6 to 10 or more in some sections for the valley, compared to hard rock, for the period range 0.2 to 3.0 seconds[1]. The effects noted are about a factor of 1.5 to 2 greater than have been observed in Los Angeles (Rogers and others, 1985), but are comparable to amplifications observed in the damaged zone of Mexico City (Singh, and others, 1988). The implication of such large site factors is that an earthquake of a given size at any given distance is likely to be more destructive in the Salt Lake area than in, say, the Los Angeles area.

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1. Considerable controversy continues in the scientific and engineering communities concerning the response of alluvium under conditions of strong shaking such as occurs in the near-field of a large earthquake. The amplification factors that are quoted for Salt Lake Valley are based on the measurements of distant Nevada Test Site underground nuclear tests, and strict application of these measurements to predict the response of alluvium under conditions of strong earthquake shaking represents an extrapolation. This extrapolation was shown, however, to approximate the measured response of alluvium in Los Angeles during the San Fernando earthquake (Rogers and others, 1985). The alluvium site-response issue continues to be discussed in connection with two questions, 1) is ground shaking greater on alluvium compared to rock at the same distance from a fault rupture; 2) are peak accelerations greater on alluvium compared to rock, all else equal? These questions are fundamentally related to how the alluvium shear velocity and attenuation parameters change under strong shaking. Clearly, low-amplitude site response factors cannot be applied to all levels of rock motion to estimate corresponding levels on alluvium. At some level of ground shaking and for some ground motion periods, the non-linear behavior of alluvium acts to limit the upper level of shaking. Nonetheless, the reader should be aware that large damaging levels of ground shaking are sustainable on some types of alluvium, as demonstrated clearly in the 1985 Mexico City and Chile events. In Chile, peak accelerations at several sites underlain by alluvium reached levels in the 0.6-0.7g range while the levels at sites underlain by rock at equivalent distances from the fault reached levels of only 0.15-0.25g. Continued research is required to fully understand the response characteristics of alluvium under strong ground shaking conditions, the situations under which site factors determined from low-level motions apply, and the specific behavior of the types of alluvium found along the Wasatch Front. At present, there is no information which would prevent us from erring on the side of safety, that is, that large low-amplitude site response measurements indicate the potential for relatively high ground shaking values, particularly for taller buildings and earthquakes that are some distance from the site under consideration.

2.3.2 Generally, individual buildings respond to narrow ranges of ground-motion periods (spectral ground motions) in a manner that is strongly dependent on building height; a general rule of thumb is that the period to which a building is most sensitive, i.e., its fundamental period, is equal to the number of stories divided by 10. For example, a 9-story building would have a fundamental period of about 0.9 seconds and be most sensitive to damage from ground motions of about 0.9 seconds. Thus, the effects noted by Hays and King, and others (see section 2.3.1) would have the greatest effect on structures with heights between 2 to 30 stories.

2.3.3 Moderate-to-large earthquakes at some distance could also cause more damage to high-rise structures located on deep sediment sites than might be expected from our extensive California experience. In particular, because of the nature of geologic site conditions in Salt Lake Valley, the ground-shaking hazard to high-rise structures sited over deep and soft valley sediments (fine sand and lake-clay deposits) are likely to be enhanced compared to the hazard at sites underlain by coarse sand and gravel, especially for distant earthquakes. For distant earthquakes, the ground motion levels that occur at the soft sediment sites are expected to be 6 to 10 times greater than at rock sites, for periods greater than about 0.2 s. For this reason, high-rise structures constructed on soft deep valley sediments may require special design to accommodate exceptionally large expected ground motions.

2.4 Losses from Ground Shaking and Other Effects (E.V. Leyendecker, S.T. Algermissen, L.M. Highland, D. Mabey, A.M. Rogers, C.M. Taylor, L. Reaveley)

2.4.1 **Effect of Magnitude and Location of Rupture on Economic Loss.**—North-central Utah should expect direct economic losses to reach $4.5 to $5.5 billion for a magnitude 7.5 earthquake occurring on the Wasatch fault zone. Losses in the four-county area of Davis, Salt Lake, Utah, and Weber counties could be as large as 23 percent of the $23.7 billion building inventory due to the effects of ground shaking and fault rupture. Losses in the same four counties range from $2.3 billion to $4.0 billion for a magnitude 6.5 earthquake and $830 million to $1.9 billion for a magnitude 5.5 earthquake. These estimates of losses due to ground shaking and fault rupture in the immediate vicinity of the Wasatch Fault have been made by Algermissen and others (1988) for a series of simulated earthquakes treated both as scenario (deterministic) and probabilistic. The scenario studies included earthquakes of different magnitudes occurring one at a time on the Provo, Salt Lake, or Weber segments of the Wasatch fault and an earthquake on a hypothetical fault west of Salt Lake City. The smallest losses result from rupture on the Provo segment while the largest losses result from rupture on the Salt Lake segment.

2.4.2 **Effect of Construction Type.**—Buildings and other structures that do not consider modern design requirements appropriate to the hazard can contribute greatly to the losses in an area. Unreinforced masonry buildings are particularly vulnerable to ground shaking. These are a large percentage of the building inventory in the four-county study area and contribute significantly to the losses. Other structural types likely to experience a large percentage of loss include reinforced concrete frame construction that has not been designed to resist earthquake ground motion.

2.4.3 **Effect on Deaths and Injuries.**—Rogers and others (1976) included estimates of deaths and injuries in a study of earthquake losses in the same four counties included in the economic loss study. Analysis of the events indicates that under the worst condition as many as 2,300 people would die, and 9,000 additional persons would suffer injuries requiring hospitalization or immediate medical treatment. The number of deaths could be as high as 14,000 if deaths from dam failure are included in the casualty total.

2.4.4 **Effect on State-Owned Buildings.**—As a result of the geographical concentration of the wealth of State-owned buildings, and of the limited seismic resistance of many of them, losses in a major Wasatch fault earthquake could easily reach 30 or even 40 percent of replacement value (Taylor and others, 1986).

2.4.5 **Effect on Lifelines.**—Liquefaction-induced ground failure along with other localized effects are likely to disrupt Wasatch Front water and natural gas systems. Except for the natural gas systems in Utah and Weber Counties, no service should be expected following a major localized earthquake. Thousands of water pipe breaks may occur and hundreds of natural gas main breaks may occur (Taylor and others, 1988; Highland, 1986). Rogers and others (1976) also examined effects on different types of lifelines. They concluded that there would be at least temporary disruption to the transportation systems—including highways, bridges, airports, and railways. There would be collapses of some structures due, in part, to earthquake resistance not being included in their design requirements.

2.4.6 **Effect on Water Impoundment Systems.**—There may be losses relating to the Great Salt Lake and Utah Lakes from a major earthquake. These losses will vary depending on factors such as lake elevations and tectonic deformation. The lake beds are areas of high liquefaction potential and dikes constructed on the lake beds are likely to be damaged and may fail in a major shaking event. Seiches may cause the Great Salt Lake and Utah Lake to oscillate for many hours, temporarily raising and lowering the water level, compounding the problem. Dike failure and or tectonic deformation of the lake beds could result in rapid inundation of some developed areas adjacent to the lakes with large losses of life and property. The study by Rogers and others (1976) examined possible dam failures. They concluded that there would be at least one dam failure and examined its effects.

2.4.7 **Other Loss Effects.**—Except for fault rupture, economic losses from liquefaction, landslides, and other ground failures have not been estimated in the Algermissen and others (1988) study. These would only increase the losses. Additional losses could be expected from factors such as fire and flooding due to dam or other water impoundment failure.

3.0 HAZARDS ASSOCIATED WITH GROUND FAILURE

3.1 Ground Failure Hazard (T.L. Youd, L. Anderson, C. Taylor)

3.1.1 Sediments susceptible to liquefaction lie beneath many areas in the lower parts of the valleys along the Wasatch Front. Large earthquakes (magnitude greater than 7) are likely to trigger liquefaction and destructive ground failures in many of these sediments. Small to moderate earthquakes (magnitude 5 to 7) are likely to trigger liquefaction locally with less severe effects.

3.1.2 The most common consequences of liquefaction along the Wasatch Front are expected to be lateral spreads. These ground failures, which occur on gentle slopes, would cause ground displacements of up to several feet along with fracture of buildings, roads, and other surface works located on the unstable ground. Pipelines and other buried facilities passing through the spreads would likely be broken or severed. Displacements capable of causing damage in the most susceptible sediment might be expected locally on the average of once in a hundred years. Larger and more widespread displacements would be associated with the more infrequent large earthquakes.

3.1.3 Major damage from rock falls, rock slides, and other slope instability should be expected on steep slopes such as within canyons and along mountain fronts. For earthquakes larger than magnitude 6.5, these failures would be distributed over a rather wide area. Smaller earthquakes could cause similar failures, but only within a few miles of the earthquake source. Facilities most commonly disrupted by these types of failures are lifelines such as pipelines, powerlines, and roads.

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Youngs, R.R., Swan, F.H., Power, M.S., Schwartz, D.P., and Green, R.K., 1987, Probabilistic analysis of earthquake ground shaking hazard along the Wasatch Front, Utah, *in* Gori, P.L., and Hays, W.W., eds., Assessment of regional earthquake hazards and risk along the Wasatch front, Utah: U.S. Geological Survey Open-File Report 87-585, p. M1-M110 [edited version in press, 1991, U.S. Geological Survey Professional Paper 1500].

APPENDIX B

UTAH'S EARTHQUAKE THREAT

THE UNIVERSITY OF UTAH

Seismograph Stations
705 W.C. Browning Building
Salt Lake City, Utah 84112
(801) 581-6274

As early as 1883, the eminent geologist G.K. Gilbert recognized and warned of the serious earthquake threat posed by the Wasatch fault and other active faults in Utah despite the absence up to that time of any large earthquakes in the region since settlement by Mormon pioneers in 1847.

The Wasatch Front area is a classic example of a seismically active region having only moderate historical seismicity but high catastrophic potential from future large earthquakes. Devastation caused by the magnitude 6.9 earthquake in Armenia on December 7, 1988, gives a real-world lesson for such situations. The high death toll of at least 30,000 people in the Armenian earthquake, due primarily to the collapse of modern buildings, emphasizes the price for not heeding the threat of infrequent large earthquakes. According to Peter Yanev (an American earthquake engineering specialist), "Rarely has the importance of systematic risk identification and proper seismic design and construction in earthquake-prone areas been more apparent (than in the Armenian earthquake)" (*EPRI Journal*, June 1989, p. 24).

Seismologists, geologists, and engineers are in fundamental agreement about technical details of the earthquake threat in Utah—where, how big, how often, and what's going to happen. That consensus, summarized below, was arrived at as part of a special five-year focus (1983-1988) on the Wasatch Front region under the National Earthquake Hazards Reduction Program.

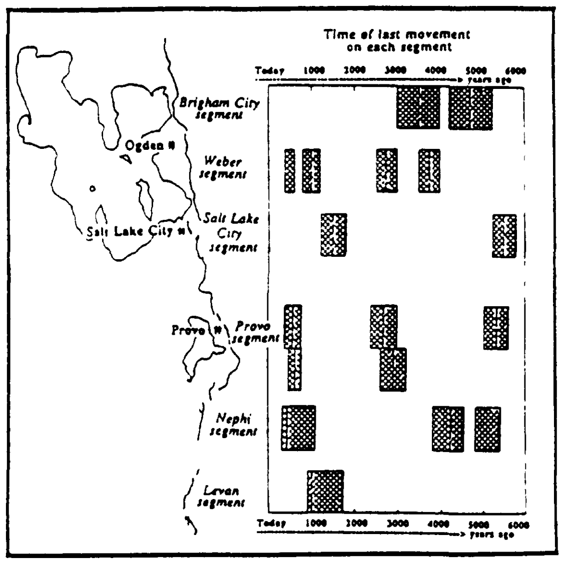
• When and where do large earthquakes occur in Utah?

— Large earthquakes (magnitude 6.5 to 7.5) can occur on any of several active segments of the Wasatch fault between Brigham City and Levan (see Figure on right). Such earthquakes can also occur on many other recognized active faults in Utah.

— During the past 6,000 years, large earthquakes have occurred on the Wasatch fault on the average of once every 400 years, somewhere along the fault's central active portion between Brigham City and Levan.

— The chance of a large earthquake in the Wasatch Front region during the next 50 years is about 1 in 5.

Prehistoric Earthquakes on Segments of the Wasatch Fault



Timing of large prehistoric earthquakes on the central part of the Wasatch fault during the past 6,000 years. Note the irregular pattern of occurrence. Heavy dashed lines are best estimates of faulting and cross-hachure pattern represents likely limits for timing as determined by radiocarbon and thermoluminescence dating. Adapted from *Segmentation models and Holocene movement history of the Wasatch fault zone, Utah*, by Machette and others, 1989, U.S. Geological Survey Open File Report 89-315, pp. 229-245.

• What would happen if a magnitude 7.5 earthquake occurs along the Wasatch fault?

— Future large earthquakes will break segments of the fault about 20 - 40 miles long and produce displacements at the surface of up to 10 - 20 feet.

— Strong ground shaking could produce considerable damage up to nearly 50 miles from the earthquake.

—The strong ground shaking may be amplified by factors up to 10 or more on valley fill compared to hard rock.

—Also possible are soil liquefaction, landslides, rock falls, and broad permanent tilting of valley floors possibly causing the Great Salt Lake or Utah Lake to inundate parts of Salt Lake City or Provo.

• How much damage would be caused by a large earthquake on the Wasatch Front?

— If the earthquake were to occur on a central part of the Wasatch fault, Utah should expect damage to buildings to exceed $4.5 billion in Davis, Salt Lake, Utah and Weber counties. This may only represent 20% of the total economic loss.

— A 1976 study by the U.S. Geological Survey for a worst case earthquake on the central Wasatch fault estimated 2,300 casualties (assuming no dam failures), 9,000 injured and 30,000 homeless. The experience of the 1988 Armenian earthquake—and engineering judgment about the collapse potential of many Wasatch Front structures—suggests the 1976 fatality estimate is low.

— Unreinforced masonry buildings (for example, brick homes built before 1960) are particularly vulnerable to ground shaking and are expected to account for 75% of the building losses.

— Surface faulting and ground failures due to shaking during a large earthquake will cause major disruption of lifelines (utilities, water, sewer), transportation systems (highways, bridges, airports, railways), and communication systems.

"It is useless to ask when this [earthquake] disaster will occur. Our occupation of the country has been too brief for us to learn how fast the Wasatch grows; and, indeed, it is only by such disasters that we can learn. By the time experience has taught us this, Salt Lake City will have been shaken down..."

– G. K. Gilbert, 1883

"Whatever the earthquake danger may be, it is a thing to be dealt with on the ground by skillful engineering, not avoided by flight...."

– G. K. Gilbert, ca. 1906

• Do we need to worry only about large earthquakes causing damage?

— No. A moderate-sized earthquake that occurs under an urbanized area can cause major damage.

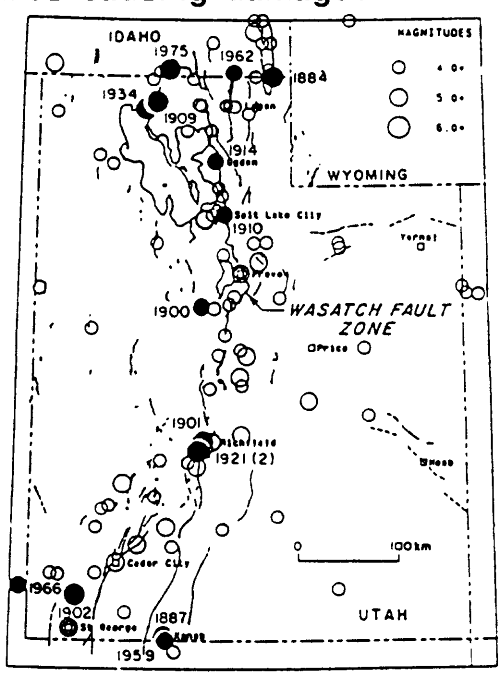
— Magnitude 5.5 - 6.5 earthquakes occur somewhere in Utah on the average of once every 7 years.

— Estimates of damage from a "direct hit" to one of the Wasatch Front's major metropolitan areas reach $2.3 billion for a magnitude 6.5 earthquake, and more than $830 million for a magnitude 5.5 earthquake.

— Since 1850, at least 15 independent earthquakes of magnitude 5.5 and larger have occurred in the Utah region (see Figure at right).

Recent magnitude 5.0 and larger earthquakes in the Utah region include:

Local Date	Magnitude	Location
Jan. 29, 1989	5.4	16 miles SE of Salina
Aug.14, 1988	5.3	Central Emery County
Mar. 27, 1975	6.0	Pocatello Valley (Utah - Idaho border)
Oct. 14, 1967	5.2	Marysvale
Aug. 16, 1966	5.6	Utah-Nevada Border
Sep. 5, 1962	5.2	Salt Lake Valley
Aug.30, 1962	5.7	Cache Valley



Epicenter map of all earthquakes of magnitude 4.0 and larger, (excluding foreshocks and aftershocks), in the Utah region from 1850 through 1989. Earthquakes of estimated magnitude 5.5 and greater are indicated by solid circles and labeled with date. Adapted from *Observational seismology and the evaluation of earthquake hazards and risk in the Wasatch Front area, Utah*, by Arabasz and others, 1989, U.S. Geological Survey Professional Paper.

• When were the largest historical earthquakes in Utah?

Since settlement in 1847, Utah's largest earthquakes were the 1934 Hansel Valley earthquake, north of the Great Salt Lake, magnitude 6.6, and the 1901 earthquake near the town of Richfield, estimated magnitude 6.5

• How often do earthquakes occur in Utah?

About 700 earthquakes (including aftershocks) are located in the Utah region each year. Approximately 2% of the earthquakes are felt. An average of about 13 earthquakes of magnitude 3.0 or larger occur in the region every year. Earthquakes can occur anywhere in the state of Utah.

• How many earthquakes occur in the Wasatch Front region?

About 500 earthquakes are located in the Wasatch Front region each year. About 60% of the earthquakes of magnitude 3.0 and larger in Utah occur in the Wasatch Front region.

• When was the last earthquake?

Worldwide: In the last minute, somewhere in the world.
Utah: Within the past 24 hours, somewhere in the state. (The last *large* earthquake in Utah occurred on the Wasatch fault north of Nephi about 400 years ago.)

• When were seismographs first installed in Utah?

In 1907, by James Talmage at the University of Utah. A skeletal statewide network began in 1962. Modern seismographic surveillance in the Wasatch Front began in 1974. Computerized recording of earthquake data began in 1981.

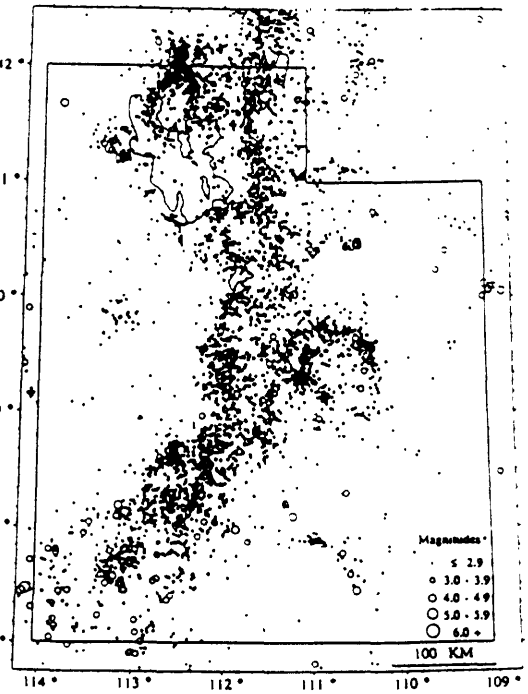
• Do earthquakes occur only on visible faults?

No. Many of the active faults in Utah are deep below the earth's surface, and are not visible to us.

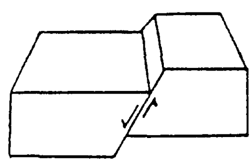
• Is the Wasatch fault the same type of fault as the San Andreas fault in California?

No. The San Andreas fault slips horizontally with little vertical movement. This is called a strike-slip fault. The Wasatch fault slips in a primarily vertical direction, with the mountains rising relative to the valley floor. The Wasatch fault is a so-called normal fault. All earthquakes produce both vertical and horizontal ground shaking. Usually the horizontal shaking is more energetic and more damaging because structures generally resist vertical loads, like gravity, more easily.

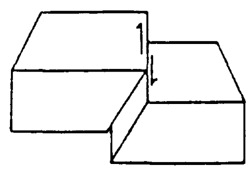
Seismicity of Utah



Each dot represents one earthquake located by the University of Utah Seismograph Stations from July 1962 through December 1989 (11,285 earthquakes).



Normal fault
(Wasatch fault type)



Strike-slip fault
(San Andreas fault type)

General Earthquake Information

• What is an earthquake?

A trembling or shaking of the ground caused by the sudden release of energy stored in the rocks below the surface, radiating from a fault along which movement has just taken place.

• How long do earthquakes last?

Generally, only seconds. Strong ground shaking during a moderate to large earthquake typically lasts about 10 to 30 seconds. Readjustments in the earth cause more earthquakes (aftershocks) that can occur intermittently for weeks or months.

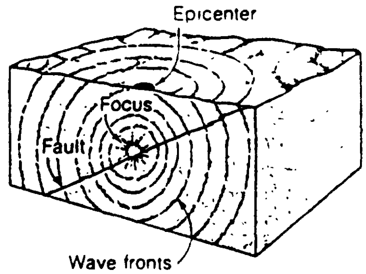


Diagram showing the focus and epicenter of an earthquake. The focus is the site of initial slip on the fault. The epicenter is the point on the surface above the focus. Also shown are seismic waves radiating from the focus.

• Is there an '*earthquake season*' or '*earthquake weather*'?

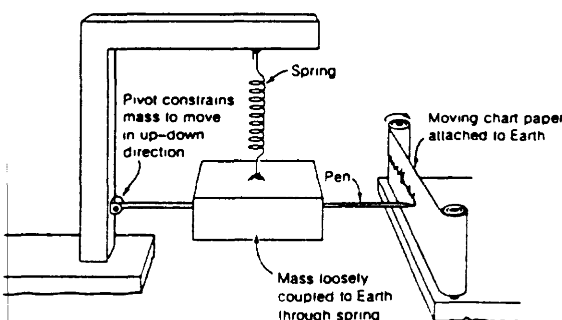
No. Earthquakes can occur at any time of the year and at any time of the day or night. Earthquakes occur under all weather conditions, sunny, wet, hot, or cold–without special tendency.

• Where is the safest place to be in an earthquake?

In an open field, where nothing can fall on you. Earthquakes do not injure or kill people; buildings and falling objects do. If you are indoors, when you feel the ground start to shake, take cover immediately under a table or sturdy piece of furniture, placing a barrier between falling objects and yourself. Do not attempt to use the stairs or an elevator or run out of the building.

• Will the ground open up during an earthquake?

The ground does not open up and swallow people (a commonly feared myth). Open ground cracks may form during an earthquake–related, for example, to landsliding or ground slumping. But such fissures are open gaps (they don't "swallow") that a person could stand in.



Cartoon depicting a seismograph that records vertical ground motion.

• What is a seismometer, seismograph, and a seismogram?

A seismometer is an sensor placed in the ground to detect vibrations of the earth. A seismograph is an instrument that records these vibrations. A seismogram is the recording (usually paper or film) of the earth's vibrations made by a seismograph.

• When was the seismograph invented?

In 1880. The earliest seismographs in the U.S. were installed in 1887, in California. (In 132 A.D. a Chinese scholar, Chang Heng, made a mechanical device to detect the first main impulse of ground shaking.)

• What is the Richter Scale?

A scale for determining the size of an earthquake from the recording of earthquake waves made on a seismograph. The maximum height of the visible recording is adjusted for the distance from the instrument to the earthquake. This is not a physical scale (in other words, one cannot look at or hold the "Richter Scale"). Each 1-unit increase in the Richter Scale roughly corresponds to a 30-fold increase in energy release and a 10-fold increase in ground motion at any site.

Magnitude	Energy equivalence
-2	100 watt light bulb left on for a week
-1	Smallest earthquake detected at Parkfield, CA
0	Seismic waves from one pound of explosives
1	A two-ton truck traveling 75 miles per hour
2	
3	Smallest earthquakes commonly felt
4	Seismic waves from 1,000 tons of explosives
5	
6	
7	1989 Loma Prieta ,CA earthquake (magnitude 7.1)
8	1906 San Francisco earthquake (magnitude 8.3)
9	Largest recorded earthquake (magnitude 8.9)

The Richter magnitude is the number generally reported in the press, and in principle the value should be the same at all recording locations (though natural variations and the use of diverse scales may lead to reported numbers that slightly differ). Due to the earth's physical limitations, the largest earthquakes have Richter magnitudes in the upper 8 range.

• Do many small earthquakes prevent larger earthquakes?

No. Observed numbers of small earthquakes are too few to equal the amount of energy released in one large earthquake. (It would take roughly 24 million earthquakes of magnitude 2 to release the same energy as one earthquake of magnitude 7.)

• Can we predict earthquakes?

Magnitude	Energy released (millions of ergs)
-2	600
-1	20000
0	600000
1	20000000
2	600000000
3	20000000000
4	600000000000
5	20000000000000
6	600000000000000
7	20000000000000000
8	600000000000000000
9	20000000000000000000

No. We cannot predict the precise time, location, and size of earthquakes in the U.S. (except in special study areas, such as Parkfield, CA). In order to predict earthquakes there has to be an adequate history of repeated earthquake cycles and/or extraordinary instrumental observations. Long-term forecasts (on scales of years or decades) are becoming common for well-studied earthquake zones. The Chinese have correctly predicted some earthquakes, evacuated cities and saved lives. They have also had large earthquakes occur with no predictions and have predicted earthquakes that never occurred.

• What is liquefaction?

Water-saturated sands, silts, and other very loosely compacted soils, when subjected to earthquake motion, may be rearranged, thereby losing their supporting strength. When this occurs, buildings may partly sink into the ground and sand and silts may come to the surface to form sand flows. In effect, the soils behave as dense fluids when liquified.

Need more information? Contact the following:

For questions about earthquake preparedness:

Utah Division of Comprehensive Emergency Management
1543 Sunnyside Avenue
Salt Lake City, UT 84108
(801) 584-8370

For questions about geology, faulting, and natural hazards in Utah:

Utah Geological and Mineral Survey
606 Black Hawk Way
Salt Lake City, UT 84108
(801) 581-6831

For questions about earthquakes:

University of Utah Seismograph Stations
705 William Browning Building
Salt Lake City, UT 84112
(801) 581-6274

For specific geologic information in Salt Lake County:

Salt Lake County Geologist
Salt Lake County Planning
2001 South State Street, Room N3700
Salt Lake City, UT 84190-4200
(801) 468-2061

For general geologic information:

U.S. Geological Survey
Public Inquiries Office
125 South State Street
8th Floor, Federal Building
Salt Lake City, UT 84138
(801) 524-5652

Appendix F

Memo, dated June 10, 1991, presented to NEPEC by Langbein outlining a proposal to revise the Parkfield earthquake prediction scenario.

OFFICE OF EARTHQUAKES, VOLCANOES, AND ENGINEERING
Branch of Tectonophysics
345 Middlefield Road, MS/977
Menlo Park, CA 94025

June 10, 1991

Memorandum

To: Parkfield Working Group

From: John Langbein

Subject: Proposed Revision to the Scenarios Document, OF 87–192

From the meeting in mid-April 1991 between representatives from USGS, OES and CDMG, it became apparent that some revisions are needed to the Parkfield, CA., Earthquake Prediction Scenarios and Response Plans. In particular, the items 5 and 6 of the attached memo from Andy Michael, dated May 22, need to be addressed. The current document implies that alerts generated by signals detected by instruments that measure deformation increase the likelihood of the anticipated Parkfield Earthquake (Table on page 2, OF 87–192). We have no data to support this conjecture. Only with the years of seismicity data collected in Parkfield and elsewhere in the State can we compute the probability that a given earthquake is a foreshock to a larger earthquake. We must state explicitly that the probabilities are for earthquakes only, and that the anomalies detected by the other networks mean "something else". Here, I propose that the "something else" is that the signal exceeds the background noise at a given confidence level. For example, a D-status is due to signals which exceed the background noise by a factor of 2, implying a confidence of 95%. The attachment to this memo explores a method to estimate the background noise, to determine the size of the signals needed to exceed the noise, and to estimate the confidence that these signals are above the background noise levels. The goal here is to define anomalous signals under a unified scheme. The current Scenarios document is hap-hazard with each investigator defining the signal level that is "anomalous". Finally, with a scheme that is common across all instrument types, we can define signal-to-noise ratios that constitute higher level alerts for all the instrument types.

The second issue concerns the fall-out of the B-level alert in March that was likely caused by a rainfall induced "creep event". The easy way to eliminate future alerts owing to the effects of rainfall for which the recorded signals are large and unprecedented is to add the phrase "with confirming signals of tectonic origin on another network". The question raised is the definition of "tectonic origin". At one end of the spectrum of definitions, signals of "tectonic origin" must have a long spatial wavelength of more than several kilometers and must be consistent with a physical model, which for Parkfield, means the model should involve fault slip on a known fault. The event last March satisfies parts of this definition in that the apparent slip was detected at 2 creepmeters separated by 3 km, but other nearby instruments located off the fault did not detect any significant changes. Although the creep is consistent with slip on the San Andreas fault, models representing the data from the creepmeters and the other strain instruments implied that slip was very shallow, less than 0.1 km depth. However, to quickly estimate the tectonic significance is not necessarily a realistic option given the possible time constraints in the *Prediction* effort.

At the other end of the spectrum of defining "tectonic origin" is the residual approach. If the signal cannot be explained or associated with a meteorologic event, localized soil creep, cultural signals, hydrological effect, or a spurious instrumental malfunction, then the signal must be of tectonic origin. For an experiment like the Parkfield prediction effort, the second definition is perhaps the more appropriate one since we really do not know for sure the type of signal that could be a precursor to the next large earthquake at Parkfield. However, for the high level alerts, I would like to see the confirming signals be significant at the 90% to 95% confidence level and that the confirming instrument be from a second network and be separated by 1 km or so from the instrument giving the high-level alert but still contiguous. For instance, if there is large creep event at the Middle Mountain instrument (XMM1) and a detectable signal recorded on a dilatometer south of Highway 46 at Jack Canyon, then I tend to down-rate the significance of the creep unless there is a detectable signal at one or more of the nearby instruments; Donnalee, Frolich, and Vineyard Canyon dilatometers, or the Flinge Flat or Vineyard Canyon water wells.

The use of a new set of rules raises many more issues that we must address. I'll suggest a few and add my opinions. One should read the following issues section, the attached, then re-read the issues section.

1. What constitutes a representative section of the data for the actual calculation of noise? My feeling is that for each instrument, the actual noise calculations should be performed on sections that have been "cleaned" using standard algorithms that remove the known signals such as the Earth tide, barometric effects, long–term hydrology, grout curing, long–term strain accumulation (a linear trend), seasonal fluctuations, and where possible, the effects from rainfall. Furthermore, the record should not contain any large, obvious tectonic events. However, the record will most likely contain signals from small tectonic events that are almost indistinguishable from noise. Such is the background from which we must quickly distinguish a significant signal from the usual noise.

2. With the use of a matched filter, it becomes possible to define the higher alert status for all the instrument networks. Currently, we have defined A and B levels for creep. If the results shown in the attached table provide a guide, we could define D at a signal to noise ratio of 2, C as 5 *to* 1, B as 10 *to* 1, and A as 20 *to* 1. Confirming signals for the A and B levels must be at the 90% confidence level (or a signal to noise ratio of 1.65) on an instrument from a second network and not be in the same instrument cluster. Furthermore, the second instrument must be contiguous with the first.

3. Should we keep the present set of combination rules to obtain higher level alerts? My feeling is yes, we should keep the current rules for now. However, it may be possible to use an extension of the scheme outlined below to modify the combination rules. For instance, all the instruments (strain, creep, two-color, water wells, and magnetometers) could be considered as one network, and the seismic network is considered the second. Hence, the combination rules come into effect with both seismicity and ground deformation.

4. The use of matched filters allows us to identify long–term rate changes which may occur over periods of weeks, months, or even years. The question is whether we should factor into the alert criteria these potential rate changes that occur over long periods. If these long-term changes do occur, then it will take a correspondingly long time to detect when the rate has dropped back to its nominal level. This is not consistent with the current scenarios document whereby background is achieved in 3 days after the onset of an alert. Since we are concentrating on short-term prediction, I suggest that we should ignore significant signals with periods greater than 2 weeks for formal alert status, but we should note the presence of these signals in the Monthly reports.

Determining signals in the presence of noise

The following attempts to summarize the discussions on "matched filters" found in many text books on time-series analysis.

Most instruments that measure deformation in the earth have a power density spectrum of noise that is proportional to $1/(f^2)$ for periods between hours to years, where f denotes frequency. The possible exception is geodetic data whose spectrum is frequency independent at the "higher frequencies". To simplify the discussion and the calculation, the noise will either be proportional to $1/(f^2)$ or a combination of $1/(f^2)$ noise at long periods and flat at the short periods.

In the following the development, I will do the derivations in the frequency domain, but the actual *back-of-the-envelope* calculations can be done in the time domain.

To estimate the power spectrum of a given instrument, sections of data should be selected that have been cleaned of telemetry glitches (or survey blunders), the earth tide and barometric response removed, and any other routine "cleaning" for the effects of grout curing, hydrology, large rainfall responses, seasonal fluctuations and the long-term secular rate. These cleaning algorithms should be those that are routinely applied to the data. Furthermore, the sample data must not include "large" events that are clearly tectonic. What we are trying to measure is the background noise of the instrument (and its site) and this noise consists of the "earth" noise plus the affect of small signals that may be due to fault slip, strain accumulation, rainfall and etc.

Once the data has been cleaned, the data variance, σ^2, can be defined in terms of the power density function, $\phi(f)$, as

$$\begin{aligned}\sigma^2 &= \int_{f_\ell}^{f_n} \phi(f)df \\ &= \int_{f_\ell}^{f_n} \frac{P}{f^2}df\end{aligned} \tag{1}$$

for instruments having $1/(f^2)$ noise and

$$\sigma^2 = \int_{f_\ell}^{f_n} \frac{Z(f^2+f_o^2)}{f^2}df \tag{2}$$

for instruments having a combination of $(1/f^2)$ and frequency independence. Here, f_n, is the nyquist frequency, f_ℓ, is the inverse of the length of time of the data window, and f_o is the cross-over frequency between the frequency independent and dependent components. If the variance of the data is calculated from the residuals from the "cleaning" algorithms and the form of the power spectrum is *assumed*, then it is easy to compute the spectral constants of either P or the combination of Z and f_o. Thus for power spectrum having assumed $1/(f^2)$ noise, then

$$P = \frac{\sigma^2 f_\ell f_n}{f_n - f_\ell} \cong \sigma^2 f_\ell \text{ if } f_n >> f_\ell \tag{3}$$

For power spectra having both frequency dependent and independent components, then we characterize the data with two variances, the variance, σ_o^2 over the entire time of the record, $1/f_\ell$, and the variance of the white noise component, σ_i^2. The cross-over frequency between the frequency dependent and independent part of the spectrum, f_o, is computed from the ratio, r, between the variance of the frequency dependent part $(\sigma_o^2 - \sigma_i^2)$ and the white noise variance, σ_i^2,

$$f_o = \left[f_n f_\ell \frac{(r\, f_n + f_\ell)}{(r\, f_\ell + f_n)}\right]^{1/2} \tag{4}$$

Then the parameter, Z, can be computed as:

$$Z = \frac{\sigma_i^2}{(f_n - f_o)(1 + \frac{f_o}{f_n})} \tag{5}$$

The background noise is then defined in terms of the power density function, $\phi(f)$. To detect a known signal, $s(t)$ or its fourier transform, $S(f)$, we compute the signal to noise ratio, ρ^2:

$$\rho^2 = \int_0^\infty \frac{\mid S(f) \mid^2}{\phi(f)} df \tag{6}$$

As written, the statistics of the above equation are difficult to understand. However, by a technique of pre-whitening the data, the statistics become straight forward. For data having $1/(f^2)$ noise, whitening the data is done by taking the first time derivative. However, with data having both frequency dependent and independent components, the appropriate filter is a high-pass filter with a corner frequency of f_o. Accordingly, the signal, $s(t)$ is modified by the filter to $s'(t)$. Then, the signal to noise ratio is written as:

$$\rho^2 = \int_0^\infty \frac{\mid S'(f) \mid^2}{P} df \quad \text{or} \quad \int_0^\infty \frac{\mid S'(f) \mid^2}{Z} df = \frac{1}{P} \int_0^\infty \mid S'(f) \mid^2 df \tag{7}$$

The time domain equivalent of equation 7 is:

$$
\begin{aligned}
\rho^2 &= \frac{1}{P}\int_0^\infty (s'(t))^2 dt \\
\text{or } &\frac{1}{Z}\int_0^\infty (s'(t))^2 dt
\end{aligned} \tag{8}
$$

Perhaps the most useful signal to detect is an amplitude change of ϵ over a period of τ, or

$$
\begin{aligned}
s(t) &= \frac{\epsilon}{\tau}t && 0 < t < \tau \\
&= \epsilon && t > \tau \\
&= 0 && t < 0
\end{aligned} \tag{9}
$$

To compute the detection thresholds for a given signal type, the hypothesized signal is passed through the same filter as the data. For the data with the $1/(f^2)$ spectra, the filtered signal is simply a box-car function of duration τ and amplitude ϵ/τ. Plugging the box-car function into equation 8 and factoring the 2π from the time–derivative yields the detectable changes in ϵ as a function of desired signal-to-noise ratio and the duration of the hypothesized signal, τ,

$$\epsilon = 2\pi\rho\sqrt{P\tau} \tag{10}$$

For a times series having both frequency dependent and independent components, the calculations are a bit more complex. The filtered version of the signal becomes

$$s'(t) = \frac{\epsilon}{2\pi f_o \tau}\left[(1 - e^{-2\pi f_o t})u(t) - (1 - e^{-2\pi f_o (t-\tau)})u(t-\tau)\right] \tag{11}$$

where $u(t)$ is the heavyside function. Plugging equation 11 into equation 8 and integrating from 0 to τ yields the expression for detecting changes of ϵ over the period τ,

$$\epsilon = \frac{2\pi f_o \rho\sqrt{Z\tau}}{\left[1 - \frac{2(1-e^{-2\pi f_o \tau})}{2\pi f_o \tau} + \frac{(1-e^{-4\pi f_o \tau})}{4\pi f_o \tau}\right]} \tag{12}$$

Although $s'(t)$ exists for $t > 0$, I have chosen to integrate from 0 to τ since, operationally, that is all of the data that exists. To check the above results, consider the case where $\tau > 1/(2\pi f_o)$, equivalent to examining the data for a rate change at periods for that the data are frequency dependent. In the limit, this converges to the same time dependence as equation 10. On the other hand, for $\tau < 1/(2\pi f_o)$, the result of expanding the exponential function into a 3^{rd} order polynomial, yields an inverse relation to τ;

$$\epsilon = \rho\sqrt{\frac{3Z}{\tau}} \tag{13}$$

which has the same dependence on averaging interval, τ, as if the background noise were strictly white noise.

The three plots shown show the sensitivities of various instrument types in detecting rate-changes. For a "generic" dilatometer, I have assumed that the power density spectrum is proportional to $1/f^2$, and that the variance of the data integrating over a 100 day period is (34 nanostrain)^2. For creep and two-color data, I have assumed that the long-period components have a $1/(f^2)$ power density spectrum, and are frequency independent at the higher frequencies. For the two-color data, I assumed that the variance of the frequency independent part of the spectra is (0.13 ppm)^2 and for the creep data, (0.1 mm)^2. The variance computed for a "generic" set of two-color data is (0.18 ppm)^2 over a 2500 day interval, and for 1000 days of creepmeter data, the variance is taken as (2.1 mm)^2.

Finally, I have set the signal-to-noise parameter, $\rho = 2$, so that it corresponds to a 95% confidence interval. The results of estimating the threshold of detectibility for each "typical" instrument are shown in the attached figure. If we compare the numbers obtained here for the creepmeters and for the dilatometers with those in the Open File report, the analysis here gives similar values for the "d-level" anomalies for the specified periods in the report. The two-color threshold levels in the Open File report are sufficiently vague such that a comparison is not possible. For the present definition of "C-level" status, both $C(2)$ for creep and the dilatometer, the equivalent signal to noise ratio is 5. Finally, for creep, the present definition of a "B-level" corresponds to $\rho = 6.7$, and "A-level" corresponds to $\rho = 25$.

Minimum number of instrument sites needed for a D-level status

The next question concerns the number of instruments that are recording "anomalous" data, that are need to identify an "alert". For instance, we have 6 dilatometers at Parkfield. At any given time, the probability is 0.05 that an individual instrument is recording an anomalous signal which exceeds its background noise by a factor of 2 or more. However, since we have 6 instruments, the probability is 0.27 that one or more instruments is recording a signal which is factor of two more than the noise. So it is likely that at least one instrument from a given network would be detecting a signal which exceeds the background noise.

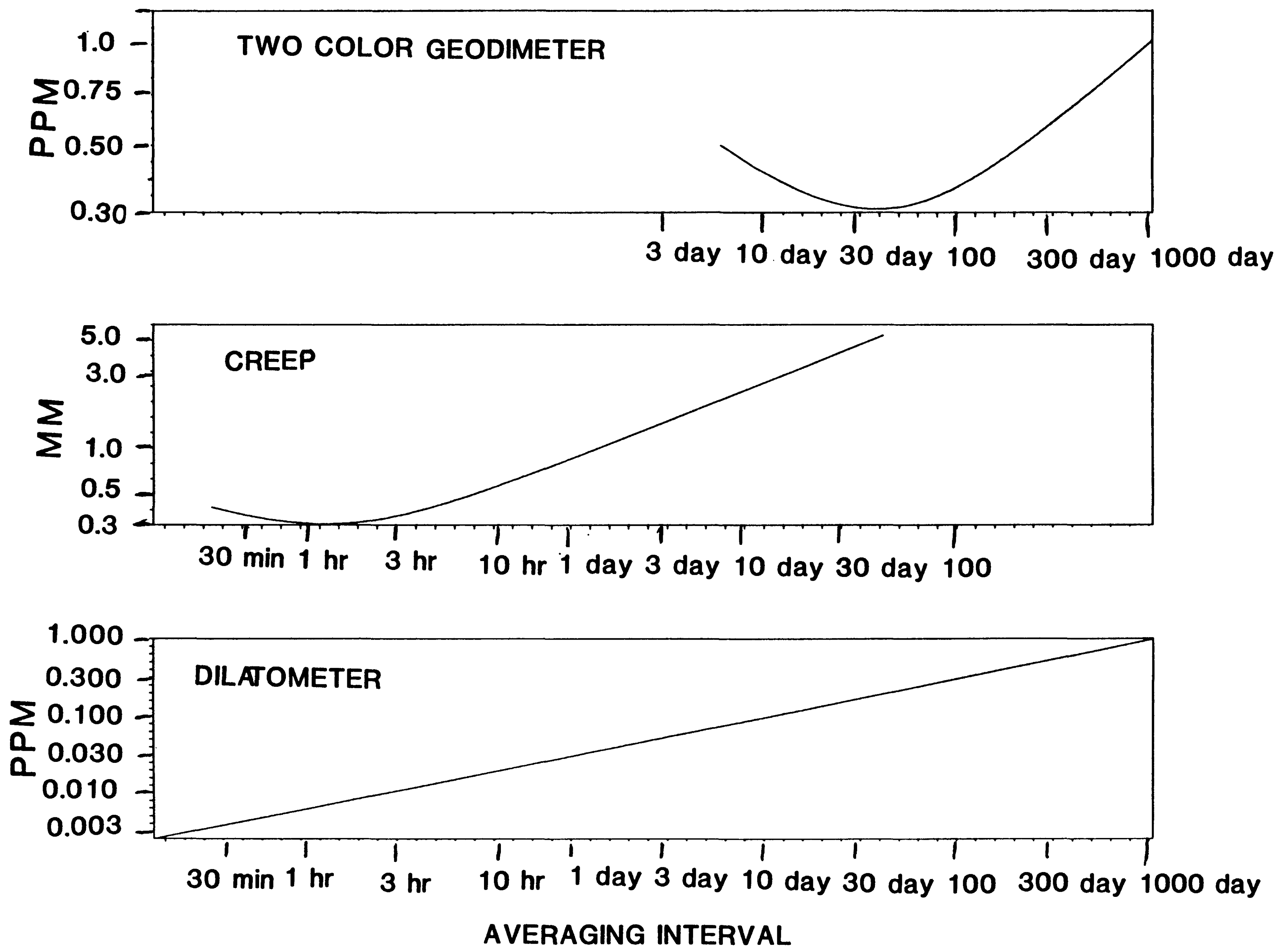
However, we can require that a second instrument is also detecting a signal. Here, the probability drops to 0.03 for two or more instruments detecting a signal which exceeds the background noise by a factor of two.

With the 17 baseline two-color network, it is likely that 1 or more baselines exceeds it background noise by a factor of two (in fact, the probability is 0.58). However, if we require that 3 or more baselines to show signal to noise ratios greater than 2.0, then the probability drops to 0.05, and for 4 baseline case, the probability drops to 0.01.

Comparison of Alert Thresholds in Open File Versus
Thresholds Calculated in This Report

Level	Open File Description	Calculated thresholds using Signal/Noise = 2. on one instrument
	CREEP	
D(1)	1 mm in 7 days	2.25 mm
D(2)	0.5 mm in 1 day, two sites	1.0 mm
D(3)	0.33–0.5 mm in 30 minutes	0.4 mm
D(4)	1.5 mm in 3 hours	0.4 mm
C(1)	0.5 mm in 1 hour, two sites	0.4 mm
C(2)	1 mm in 1 hour at Middle Mountain	need to compute 0.4 mm???
B	5 mm within 3 days	1.5 mm
A	5 mm within 10 hours	0.4 mm
	DILATOMETERS	
D(1)	0.10 ppm in 7 days at two sites	0.08 ppm
D(2)	0.10 ppm in 1 day at one site and indications at second site	0.03 ppm
C(1)	0.20 ppm in 7 days at two sites	0.08 ppm
C(2)	0.20 ppm in 1 day at one site and indications at second site	0.08 ppm

SENSITIVITY for $\rho=2$



United States Department of the Interior

GEOLOGICAL SURVEY
OFFICE OF EARTHQUAKES, VOLCANOES, AND ENGINEERING
Branch of Seismology
345 Middlefield Road - Mail Stop 977
Menlo Park, California 94025
May 22, 1991

MEMORANDUM

To: Richard Andrews (OES), Bill Bakun (USGS), Jim Davis (CDMG), Pat Jorgenson(USGS), John Langbein (USGS), Al Lindh (USGS), Tom Mullins (OES), Will Prescott (USGS)

From: Andy Michael

Subject: Changes in the USGS Parkfield Scenarios and Response Plan as per the April 11, 1991 Meeting, *revised version*

Below I detail the changes the USGS will be making in the Parkfield Scenarios and Response Plan in response to the meeting of April 11, 1991. I would appreciate your comments on these changes so that we are able to do the best possible job when revising the Parkfield Scenarios and Response Plan Open-File Report. Note that the changes under 1 and 2 were put into effect immediately after the meeting.

1. The word "Alert" will no longer be used for C through E level situations. Instead we will use the term "Status." As in, "The Parkfield Experiment is at C Status due to two M>1.5 earthquakes under Middle Mountain." The use of "Status" will be incorporated into the the Long Valley Caldera plan and the plan for the Coachella Valley segment of the San Andreas fault. This change is being made because the use of the word "Alert" overstates our level of concern that these circumstances should be considered as precursors to the Parkfield earthquake. To many of the people who learned of the situation, the word "Alert" appeared to connote a recommendation for an unwarranted level of action.

2. On the basis of experiences with C level Status conditions over a 6-year period, both the USGS and OES feel that information of on the C Status is worth of note by the members of the Parkfield working group and also to the OES earthquake program and regional managers in Regions I, II, and V, and the emergency services coordinators in the 7 counties that are part of the Parkfield area. This will exercise part of the communications system and allow scientific interchange about these events. It is not clear that more widespread, immediate notification to other groups or regions is necessary or that it enhances emergency preparedness in a meaningful way.

   Thus, at C Status the USGS will not send out any written statements about the events that lead to the change in Status or the significance of the change. Instead the USGS will notify the members of the Parkfield Working Group and verbally notify the OES Warning Center. To keep the public informed, C Statuses will be mentioned in the weekly report on Northern California seismicity. This report will include a generic statement that will clarify that the occurrence of a C status is not considered to warrant any public action (see 5). The weekly report will be sent to the county OES offices affected by the Parkfield experiment, the CA OES office and state regions I, II, and V.

3. Should any of the parties involved in the Parkfield experiment (e.g. the USGS, the California OES, the CDMG, or any of the 7 counties) become aware of media/public information regarding the "C" level Status prior to the distribution of the weekly USGS seismicity report, it is their responsibility to notify the other parties as soon as possible. On the part of the USGS we will notify the CA OES and the CDMG and CA OES will notify the counties.

   At that time, the USGS, in consultation with CA OES, will determine whether a press statement is appropriate in order to make media coverage as accurate as possible. The contents of this press statement will also be determined by consultation between the USGS and CA OES.

4. The terminology for B level will continue to be "Alert" and, following OES' notification to the Parkfield area counties, the USGS shall prepare a press statement that is to be coordinated with the OES Directory of Public Information that sets forth the circumstances that led to the B level Alert. A pre-prepared press statement for level B alerts will be drafted by the Chief Scientist for review by NEPEC and CEPEC with space to insert the information that is specific to each alert (e.g. the type of anomaly that lead to the alert and any other unusual circumstances.). OES will in turn prepare a press statement including recommended public actions and guidance for local jurisdictions. OES will provide copies of both press statements to OES regions for relay to all counties for action or information purposes, as appropriate.

5. A statement that places the different Status and Alert levels into a general context will be prepared by the USGS working group to replace the use of numerical probabilities when discussing Status and Alert levels not caused by seismicity. These context statements will become part of the USGS Scenario document.

6. Changes to the Status/Alert criteria will be considered by the Working Group and a revision of the USGS Scenario Document will be prepared for discussion by NEPEC and CEPEC their June and July 1991 meetings respectively. Among these changes will be the need for confirmation of tectonic changes from other networks for all A and B level creep alerts.