1 Introduction

I cannot help feeling that seismology will stay in the place at the center of solid earth science for many, many years to come.
The joy of being a seismologist comes to you, when you find something new about the earth's interior from the observation of seismic waves obtained on the surface, and realize that you did it without penetrating the earth or touching or examining it directly.

Keiiti Aki, presidential address to the Seismological Society of America, 1980

1.1 Introduction

This book is an introduction to seismology, the study of elastic waves or sound waves in the solid earth. Conceptually, the subject is simple. Seismic waves are generated at a *source*, which can be natural, such as an earthquake, or artificial, such as an explosion. The resulting waves propagate through the *medium*, some portion of the earth, and are recorded at a *receiver* (Fig. 1.1-1). A *seismogram*, the record of the motion of the ground at a receiver called a *seismometer*, thus contains information about both the source and the medium. This information can take several forms. The waves provide information on the location and nature of the source that generated them. If the *origin time* when the waves left the source is known, their *arrival time* at the receiver gives the *travel time* required to pass through the medium, and hence information about the speed at which they traveled, and thus the physical properties of the medium. In addition, because the amplitude and shape of the wave pulses that left the source are affected by propagation through the medium, the signals observed on seismograms provide additional information about the medium.

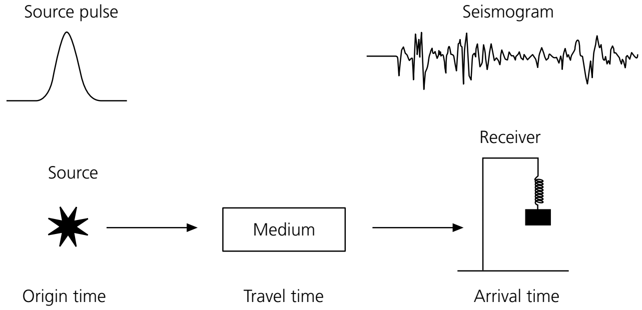


Fig. 1.1-1 Schematic geometry of a seismic experiment.

1.1.1 Overview

Before embarking on our studies, it is worth briefly outlining some of the ways in which seismology is used to study the earth, and some of the methods used. Seismology is the primary tool for the study of the earth's interior because little of the planet is accessible to direct observation. The surface can be mapped and explored, and drilling has penetrated to depths of up to 13 kilometers, though at great expense. Information about deeper depths, down to the center of the earth (approximately 6371 km), is obtained primarily from indirect methods. Seismology, the most powerful such method, is used to map the earth's interior and study the distribution of physical properties. The existence of the earth's shallow crust, deeper mantle, liquid outer core, and solid inner core are inferred from variations in seismic velocity with depth. Our ideas about their chemical compositions, including the presumed locations of changes in mineral structure due to the increase of pressure with depth, are also based on seismological data. Near the surface, seismology provides detailed crustal images that reveal information about the locations of economic resources like oil and minerals. Deeper in the earth, seismology provides the basic data for understanding earth's dynamic history and evolution, including the process of mantle convection.

Seismology is also the primary method for studies of earthquakes. Most of the information about the nature of faulting during an earthquake is determined from the resulting seismograms. These observations are useful for several purposes. Because earthquakes generally result from the motions of the plates making up the earth's lithosphere, which are the surface expression of convection within earth's mantle, knowledge of the direction and amount of motion is valuable for describing plate motions and the forces giving rise to them. Analysis of seismograms also makes it possible to investigate the physical processes that occur prior to, during, and after faulting. Such studies are helpful in assessing the societal hazards posed by earthquakes.

Our purpose here is to discuss some basic ideas about seismology and its applications. To do this, we first introduce several concepts about waves in a solid medium. We will see that a few simple but powerful ideas give a great deal of insight into how waves propagate and respond to variations in physical properties in the earth. Fortunately, most of these ideas are analogous to familiar concepts in the propagation of light and sound waves. As a result, studying the earth with seismic waves is conceptually similar to sensing the world around us using light and sound. For example, you are reading this by receiving light reflected off the paper. We see color because light has different wavelengths; the sky is blue because certain wavelengths are scattered preferentially. An even closer analogy is the use of sound waves by bats, dolphins, and submarines to "see" their surroundings. Seismology gives detailed images of earth structure, much as sound waves (ultrasound) and electromagnetic waves (X-rays) are used in medicine to study human bodies.

A familiar property of light is that it bends when traveling between materials in which its speed differs. Objects inserted into water appear crooked, because light waves travel more slowly in water than in air. Prisms and lenses use this effect, called *refraction*. This phenomenon occurs in the earth because seismic wave velocities generally increase with depth. Wave paths bend away from the vertical as they go deeper into the earth, eventually become horizontal ("bottom"), turn upward, and return to the surface (Fig. 1.1-2). The wave paths are thus used to infer the variation of seismic velocity, and hence the composition and physical properties of material, with depth in the earth.

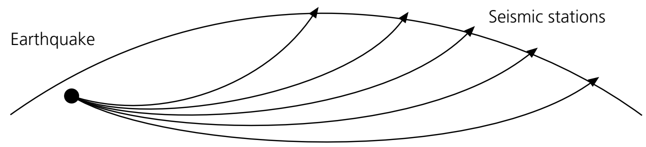
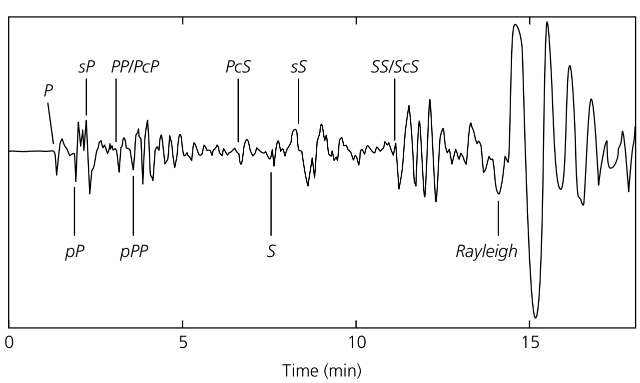


Fig. 1.1-2 Seismic ray paths in the earth, showing the effect of an increase in seismic velocity with increasing depth. The waves travel in curved paths between the earthquake and seismic stations.

Just as light waves *reflect* at a mirror, seismic waves reflect at interfaces across which physical properties change, such as the boundary between the earth's mantle and core. Because the amplitudes of the reflected and transmitted seismic waves depend on the velocities and densities of the material on either side of the boundary, analysis of seismic waves yields information on the nature of the interface. In addition to refraction and reflection, waves also undergo *diffraction*. Just as sound diffracts around the corner of a building, allowing us to hear what we cannot see, seismic waves bend around "obstacles" such as the earth's core.

The basic data for these studies are seismograms, records of the motion of the ground resulting from the arrival of refracted, reflected, and diffracted seismic waves. Seismograms incorporate precise timing, so that travel times can be determined. The seismometer's response is known, so the seismogram can be related to the actual ground motion. Because ground motion is a vector, three different components (north–south, east–west, and up–down) are typically recorded. Hence, although seismograms at first appear to be simply wiggly lines, they contain interesting and useful information.

To illustrate the use of seismology for the study of earth structure, consider a seismogram from a magnitude 6 earthquake in Colombia, recorded about 4900 kilometers away in Colorado (Fig. 1.1-3). Several seismic wave arrivals, called *phases*, are identified using a simple nomenclature that describes the path each followed from the source to the receiver.



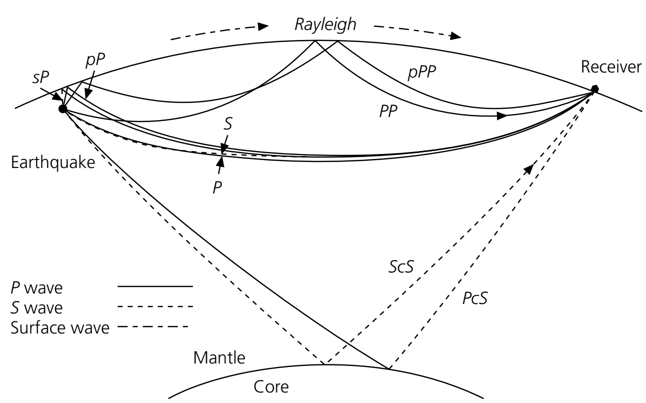


Fig. 1.1-3 *Left*: Long-period vertical component seismogram at Golden, Colorado, from an earthquake in Colombia (July 29, 1967), showing various seismic phases. The distance from earthquake to station is 44°. *Right*: Ray paths for the seismic phases labeled on the seismogram.

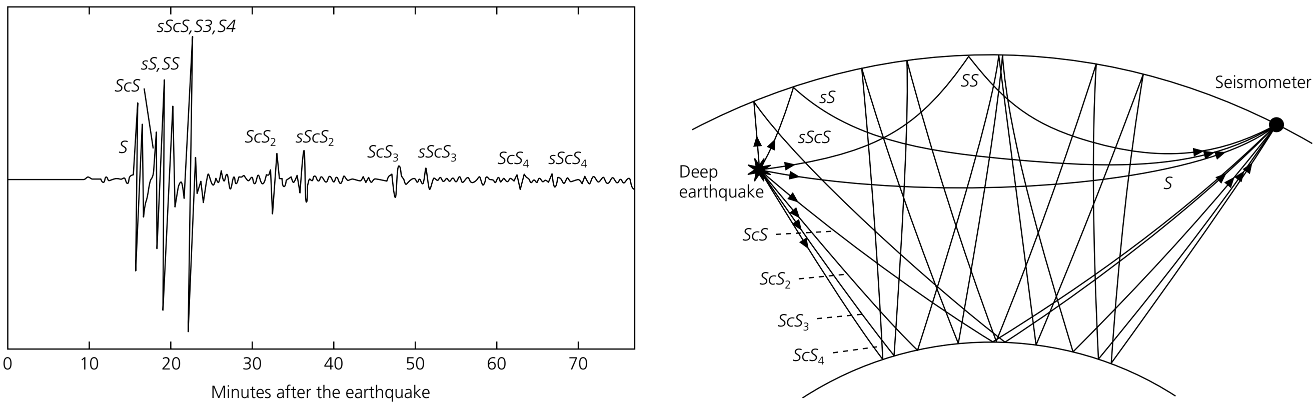


Fig. 1.1-4 Seismogram (*left*) and ray paths (*right*) for a deep focus earthquake in Tonga, recorded at Oahu (Hawaii), showing multiple core reflections.

We will see that seismic waves are divided into two types. In one type, *P* or compressional waves, material moves back and forth in the direction in which the wave propagates. In the other, *S* or shear waves, material moves at right angles to the propagation direction. *P* waves travel faster than *S* waves, so the first arriving pulse, labeled "*P*," is a *P* wave that followed a direct path from the earthquake to the seismometer.[1] Soon afterwards, a pulse labeled *pP* appears, which went upward from the earthquake, reflected off the earth's surface, and then traveled to the seismometer as a *P* wave. If the distribution of seismic velocity near the source is known, the depth of the earthquake below the earth's surface can be found from the time difference between the direct *P* and *pP* phases, because the primary differences between their ray paths are the *pP* segments that first go up to and then reflect off the surface. The phase marked *PP* is a compressional wave that went downward from the source, "bottomed," reflected at the surface, and repeated the process. Among the later arrivals on the seismogram are shear wave phases, including the direct shear wave arrival, *S*, and a shear phase *SS* that reflected off the surface, analogous to *PP*. All these phases, which traveled through the earth's interior, are known as *body waves*. The large amplitude wave train that arrives later, marked "Rayleigh," is an example of a different type of wave. Such *surface waves* propagate along paths close to the earth's surface.

Figure 1.1-4 shows a seismogram from an earthquake at a depth of 650 km in the Tonga subduction zone recorded in Hawaii. The seismometer is oriented such that all the arrivals are shear waves. In addition to *S* and *SS*, phases reflected at the core–mantle boundary appear. *ScS* went down from the source, reflected at the core–mantle boundary (hence "*c*"), and came back up to the seismometer. Its travel time gives the depth to the core if the velocity in the mantle is known. Alternatively, if the depth to the core is known, the travel time gives a vertical average of velocity with depth in the mantle. In addition, the large amplitude of these reflections constrains the contrast in physical properties between the solid rock-like lower mantle and the fluid iron outer core. Multiple reflections also occur: *ScSScS*, or ScS_2, reflects twice at the core–mantle boundary, ScS_3 reflects three times, and ScS_4 four times. Similar to the phase *SS*, the S_3 wave reflects twice off the surface, and S_4 reflects three times. By analogy to *pP*, *sScS* went upward from the source and was reflected first at the surface and then at the core–mantle boundary. Most of the multiple *SS* and *ScS* phases also have observable surface reflected phases (e.g., $sScS_2$, $sScS_3$, etc.).

These examples indicate some of the ways in which seismological observations are used to study earth structure. By collecting many such records, seismologists have compiled travel time and amplitude data for many seismic phases. Because the different phases have different paths, they provide multiple types of information about the distribution of seismic velocities, and therefore physical properties within the earth. Seismology can also be used to study the internal structure of other planets; seismometers were deployed on the lunar surface by each of the Apollo missions, and the Viking spacecraft that landed on Mars carried a seismometer.

An important use of seismology is the exploration of near-surface regions for scientific purposes or resource extraction. Figure 1.1-5 shows a schematic version of a common technique used. An artificial source at or near the surface generates seismic waves that travel downward, reflect off interfaces at depth, and are detected by seismometer arrays. The resulting data are processed using computers to enhance the arrivals corresponding to reflections and to estimate the velocity structure. Seismograms from different receivers are then displayed side by side, with the travel time increasing downward, to yield an image of the vertical structure. Reflections that match between seismograms give near-horizontal arrivals that often correspond to interfaces at depth. The vertical axis can be converted from time to depth using the estimated velocities, and reflectors

[1] The labels *P* and *S* come from the early days of seismology, when *P* stood for *primary* and *S* stood for *secondary*.

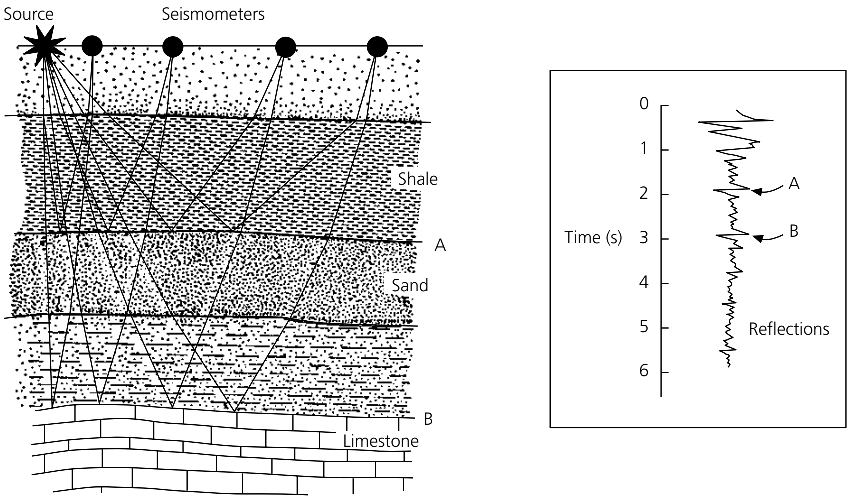


Fig. 1.1-5 Schematic example of the seismic reflection method, the basic tool of hydrocarbon exploration.

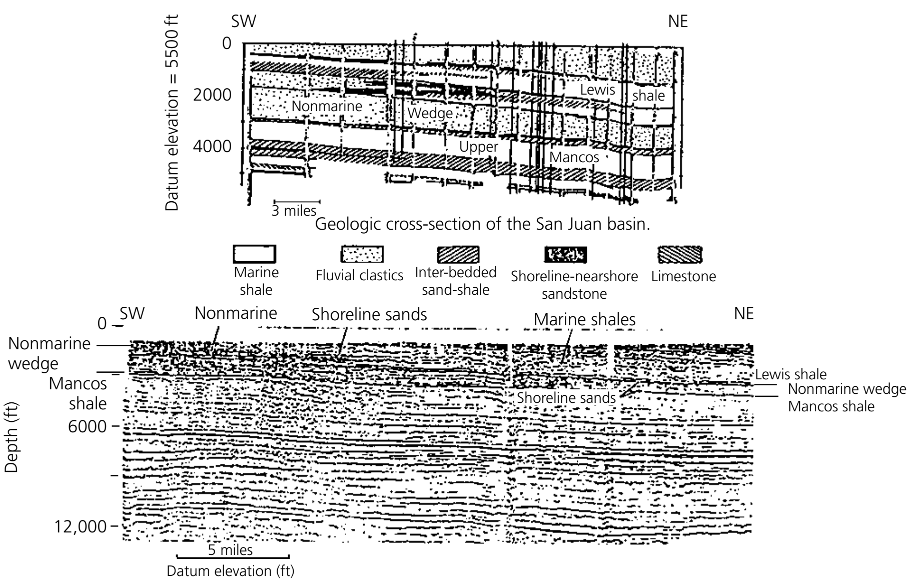


Fig. 1.1-6 Data from a reflection seismic survey across the San Juan Basin, New Mexico (*bottom*) and the resulting geological interpretation (*top*). [Sangree and Widmier, 1979. Reprinted by permission of the Society of Exploration Geophysicists.)

can be identified using geological information from the surface and drill holes (Fig. 1.1-6). Such seismic images of the subsurface provide a powerful tool for structural and stratigraphic studies. Although applications of seismology to exploration have traditionally been treated in universities as distinct from those dealing with earthquakes and the large-scale structure of the earth, this distinction is largely historical.[2] These applications draw on a common body of seismological principles, and the techniques used have considerable overlap.

Seismic sources — typically earthquakes — are also a major topic of seismological study. The location of an earthquake, known as the *focus* or *hypocenter*, is found from the arrival times of seismic waves recorded on seismometers at different sites. This location is often shown by the *epicenter*, the point on the earth's surface above the earthquake. The size of earthquakes is measured from the amplitude of the motion recorded on seismograms, and given in terms of *magnitude* or *moment*.[3] In addition, the geometry of the fault on which an earthquake

[2] This book follows this tradition and focuses on earthquakes and large-scale earth structure because of the existence of an excellent introductory literature dealing with exploration seismology and the inflexibility of university curricula.

[3] Magnitude is given as a dimensionless number measured in various ways, including the body wave magnitude m_b, surface wave magnitude M_s, and moment magnitude M_w, as discussed in Section 4.6. The seismic moment has the dimensions of energy, dyn-cm or N-m.

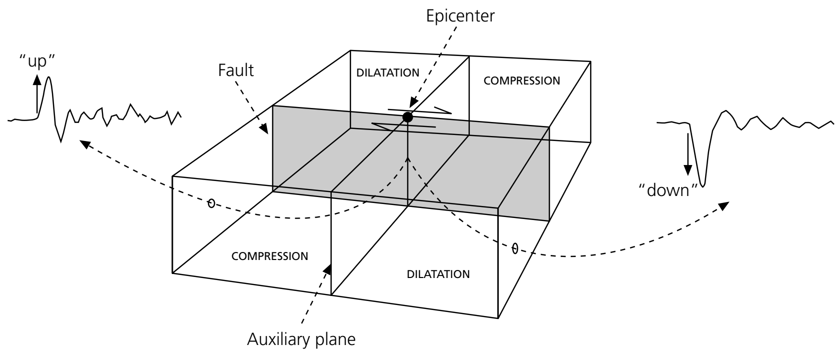


Fig. 1.1-7 First motions of seismic *P* waves observed at seismometers located in various directions about the earthquake allow the fault orientation to be determined.

occurred is inferred from the three-dimensional pattern of radiated seismic waves. Figure 1.1-7 illustrates the method used for an earthquake in which the material on one side of a vertically dipping fault moves horizontally with respect to that on the other side. This motion generates seismic waves that propagate away in all directions. In some directions the ground first moves away from the source (toward a seismic station), whereas in other directions the ground first moves toward the source (away from a receiver). The seismograms thus differ between stations. In the "toward" (called compressional) quadrants the first ground motion recorded is toward the receiver, whereas in the "away" (called dilatational) quadrants the first ground motion is away from the receiver. Because the seismic waves go down from the source, turn, and arrive at a distant seismographic station from below, the first motion is upward in a compressional quadrant and downward in a dilatational quadrant.[4] The compressional and dilatational quadrants can be identified using seismograms recorded at different azimuths around the source. The fault orientation and a surface perpendicular to it can then be found, because in these directions the first motion changes polarity. With the use of additional data we can often tell which of these surfaces was the actual fault. Given the fault orientation, the direction of motion can also be found; note that the compressional and dilatational quadrants would be interchanged if the fault had moved in the opposite direction. The pulse radiated from the earthquake also gives some information about the amount of slip that occurred, the size of the area that slipped, and the slip process.

Such observations of the location of earthquakes and the fault motion that occurred in them are among the most important data we have for understanding *plate tectonics*, the primary process shaping our planet. The earthquake analyzed in Fig. 1.1-7, for example, is like those that occur along the San Andreas fault in northern California, part of the boundary along which the Pacific plate moves northward with respect to the North American plate. The fault is visible at the earth's surface, so geological and geodetic observations also show the motion that occurs in earthquakes. In less accessible areas seismological observations provide most of the data used to identify the boundary along which motion occurs and to demonstrate its nature. This is the case for most plate boundaries, which occur in the oceans, beneath several kilometers of water. Similarly, in subduction zones, where lithospheric plates descend deep into the mantle and earthquakes can occur to depths of 660 km, direct observations are not possible, but analyses of seismograms reveal the motions and give insight into their tectonic causes.

1.1.2 Models in seismology

As summarized in the previous section, seismology provides a great deal of information about seismic sources, the structure of the earth, and the relation of earthquakes to the tectonic processes that produce them. Even so, we will see that there are major limitations on what the present seismological observations and other data tell us. For example, although we have good models of seismic velocity in the earth, we know much less about the composition of the earth and have only general ideas about the deep physical processes, such as convection, thought to be taking place. Similarly, although seismology provides a great deal of detail about the slip that occurs during an earthquake, we still have only general ideas about how earthquakes are related to tectonics, little understanding of the actual faulting process, no ability to predict earthquakes on time scales shorter than a hundred years, and only rudimentary methods to estimate earthquake hazards. This situation is typical of the earth sciences,[5] largely because of the complexity of the processes being studied and the limits of our observations. Our best response seems to be to show humility in face of the complexity of nature, recognize what we presently know

[4] These terms are not the same as compressional and shear waves; as often occurs in science, words have multiple meanings.

[5] In discussing analogous issues Sarewitz and Pielke (2000) note than even after billions of dollars spent on climate research, a senior scientist observes, "This may come as a shock to many people who assume that we do know adequately what's going on with the climate, but we don't," and the National Academy of Sciences states that deficiencies in our understanding "place serious limitations on the confidence" of climate modeling results.

and what we do not, use statistical techniques to assess what we can say with differing degrees of confidence from the data, and develop new data and techniques to do better.

In general, the approach taken is to describe complex problems with simplified models that seek to represent key elements of the process under consideration. For example, an earthquake is a complicated rupture process that occurs in a finite volume and radiates seismic energy through the real materials of the earth. As we will see in the next few chapters, we represent all aspects of this process with simple models. We treat the complex faulting process as elastic slip on an infinitely narrow surface. We further treat the rock around it as a simple elastic material, and thus describe the complex seismic wave disturbance that propagates through it, using a number of simplifications.

It is important to bear in mind that these models are only approximations to a more complicated reality. For example, although the radiated seismic energy is real (it can destroy buildings), the mathematical descriptions used to understand it are human constructs. *P* waves, *S* waves, seismic phases like *ScS*, seismic ray paths, surface waves, or the earth's normal modes are all approximations that make the radiated energy easier to conceptualize. Similarly, we model a fault as a planar slip surface and use seismological observations to characterize the slip geometry and history. However, although this process nicely replicates the seismic observations, it only approximates the actual physics of earthquake rupture.

We often use a hierarchy of different approximations, as appropriate. For example, we might first predict the approximate time when a packet of seismic energy arrives by treating it as a seismic ray, and then use a more sophisticated wave or normal mode calculation to predict its amplitude and hence learn more about the properties of the parts of the earth it traversed. Similarly, we first describe the earth as isotropic (having the same properties in all directions) and purely elastic (no seismic energy is lost to heat by friction) and then confront the deviations from these simplifications.

A similar approach is often followed when discussing the tectonic context of earthquakes. Although faults, earthquakes, volcanoes, and topography are real, we associate these with the boundaries of plates that are human approximations. We will see that the questions of when to regard a region as a plate and how to characterize its boundaries are not simple. The simplest analyses assume that plates are rigid and divided by narrow boundaries. Later, we treat the boundaries as broad zones, and eventually we confront the fact that plates are not perfectly rigid, but in fact deform internally, as shown by earthquakes that occur within them.

We often choose a type of model to represent the earth and then use seismological and other data to estimate the parameters of this model. Thus a characteristic activity of seismology, and of the earth sciences in general, is solving *inverse problems*. We start with the end result, the seismograms, and work backwards using mathematical techniques to characterize the earthquakes that generated the seismic waves and the material the waves passed through. Inverse problems are more complicated than the conceptually simpler *forward problems* in which we use the theory of seismic wave generation and propagation to predict the seismogram that would be observed for a given source and medium. Inverse problems are harder to solve for several reasons. Seismograms reflect the combined effect of the source and medium, neither of which is known exactly. There are often aspects of the inverse problem that the data are insufficient to resolve. Thus seismology and other branches of the earth sciences, to a greater extent than most other scientific disciplines, often infer a "big picture" from grossly limited and insufficient data. For example, our images of the earth from seismic waves suffer from the fact that the severely limited geographical distributions of both earthquakes and seismometers leave most of earth's interior unsampled. This situation is like a doctor examining a possible broken bone with only a few scattered bursts of x-rays from random directions.

Moreover, although the forward problem typically can be solved in a straightforward way, giving a unique solution, the inverse problem often has no unique solution. In fact, the data are generally somewhat inconsistent due to errors, so no model can exactly describe the data. Finally, the fact that solving the inverse problem yields a set of model parameters that describe the observations well does not necessarily mean that the resulting model actually reflects physical reality. This nonuniqueness reflects the logical tenet that because *a* implies *b*, *b* does not necessarily imply *a*. In fact, we often have no way of determining what the reality is. For example, we will never truly know the composition and temperature of the earth's core because we cannot go there. This limitation remains in spite of the fact that over time our models of the core have become increasingly consistent with seismological data, experimental results about materials at high pressure and temperature, and other data including inferences from meteorites about the composition of the solar system.[6]

A consequence of this approach is the need to consider issues of precision, accuracy, and uncertainty. Estimates of quantities like the magnitude or depth of an earthquake depend both on the precision, or repeatability, with which data like seismic wave arrival times and amplitudes are measured, and on the accuracy, or extent to which the resulting inferences correctly describe the earth. For example, earthquake magnitudes are simple measures of earthquake size, estimated in various ways from seismograms without accounting for effects like the geometry of the earthquake source or lateral variations in seismic velocities. Hence measurements at different sites yield various estimates, so it is of little value to argue whether an earthquake had magnitude 5.2 or 5.4. Similarly, focal depths are derived from seismic wave arrival times by assuming a velocity structure near the earthquake, which is often not well known. For

[6] Similar difficulties afflict most of the earth sciences. Field geologists will never know whether their inferences about the past history and environment of a region are correct; paleontologists will never know how realistic their models of ancient life are, etc.

example, the depth is sometimes estimated (Section 4.3.3) from half the product of the time difference between the direct P and pP phases (see Fig. 1.1-3) and the velocity. If the time difference is measured to 0.25 s, and the velocity is 8 km/s, the method of propagation of errors (Section 6.5.1) shows that the uncertainty in depth is about 1 km, so it makes little sense to report the depth to greater precision. In reality the uncertainty will be greater, because the velocity also has some uncertainty. It is important to bear in mind that assigning a single value to an earthquake depth may exceed the relevant accuracy because faulting extends over a finite area that may be large (on the order of 10 km for a magnitude 6 earthquake). Moreover, when we have alternative models with which to estimate a parameter (for example, the earthquake stress drop estimated from body waves depends on the assumed geometry of the fault), the uncertainty associated with an estimate using any particular model underestimates the uncertainty due to the fact that we do not know which model is best. It is thus useful to examine how the estimate depends on the precision of the observation, the model parameters, and the choice of models.

Seismologists generally assume that the best estimates of values and uncertainties come from studies by different investigators using multiple datasets and techniques. Ideally, studies using the same data increase precision by reducing random errors, and studies using different data and techniques increase accuracy by reducing the effect of systematic errors. For example, for the well-studied Loma Prieta earthquake, seismic moment estimates vary by about 25%, and M_s values vary by about 0.1 units.

However, statisticians have long noted the difficulties in assessing probabilities and uncertainties. Two famous examples are the *Titanic*, described as "unsinkable" (probability zero) and the space shuttle, which was lost on its twenty-fifth launch, surprisingly soon given the estimated probability of accident of 1/100,000. Other examples come from the history of measurements of physical constants, which shows that the reported uncertainties underestimate the actual errors. For example, the 27 successive measurements of the speed of light between 1875 and 1958 are shown by subsequent analysis to be consistently in error by much more than the assigned uncertainty. It appears that assessments of the formal or random uncertainty often significantly underestimate the systematic error, so the overall uncertainty is dominated by the unrecognized systematic error and thus larger than expected. As a result, measurements of a quantity often remain stable for some time, and then change by much more than the previously assumed uncertainty. One possible explanation, termed the "bandwagon effect," is the tendency to discount data that are inconsistent with previous ideas, but later prove more accurate than those included. Another effect appears to be the discarding of outliers: for example, although R. Millikan reported using all the observations in his Nobel prize-winning (1910) study of the charge of the electron, his notebooks show that he discarded 49 of 107 oil drops that appeared discordant, increasing the apparent precision of the result. Until a method is developed that excludes obviously erroneous data without discarding real disconforming evidence, making realistic uncertainty estimates will remain a challenge. Although such analyses are more difficult in the earth sciences — for example, an earthquake is a nonrepeatable experiment — they are useful to bear in mind.

This discussion brings out the fact that although we often speak of "finding" or "determining" quantities like earthquake source parameters or velocity structure, it might be better to speak of "estimating" or "inferring" these quantities. There is no harm in the common and more upbeat phrasing so long as we remember that these values reflect uncertainties due to random noise and errors of measurement (sometimes called *aleatory* uncertainty, after the Latin word for dice) and systematic (sometimes called *epistemic*) uncertainty due to our choice of model to describe the phenomenon under consideration.

Although these caveats sound worrisome, seismological models are far from useless. We can usually develop models that not only describe the data used to develop them, but to predict other data. For example, earthquake source models derived only from seismology often predict the observations made using field geology and geodesy (ground deformation), both for the specific earthquake studied and for others in the same region. Moreover, the seismological results often give useful insight that is consistent with other lines of evidence. For example, seismology, gravity, and geomagnetism all favor the earth having a dense liquid iron core chemically different from the rocky mantle. This idea is also consistent with the fact that meteorites — thought to be fragments of small planets — are divided into stony and iron classes. Hence seismologists use this modeling approach to understand the earth, while recognizing its limitations.

For several reasons, our models usually improve with time. First, the data improve in both quantity and quality. Second, new observational and analytical techniques are introduced. As a result, long-standing problems such as the velocity structure of the earth are repeatedly reassessed. Successive generations of models seek to explain additional types of data, and often contain more model parameters in the hope of better representing the earth. Using statistical tests, we find that in some cases the resulting improvements are significant, whereas in others the new model improves only slightly on earlier ones. An important point is that more complicated models can always fit data better, because they contain more free parameters, just as a set of points in the x–y plane can be better fit by a quadratic polynomial than by a straight line. Thus we can statistically test models to see whether a new model reduces the misfit to the data more than would be expected purely by chance due to the additional parameters. Another useful test is whether the new or old models do a better job of describing data that were not used in deriving either, a process called pure prediction. When new models pass these tests, we can accept them — and then look again to see which data are still not described well and try to do better.

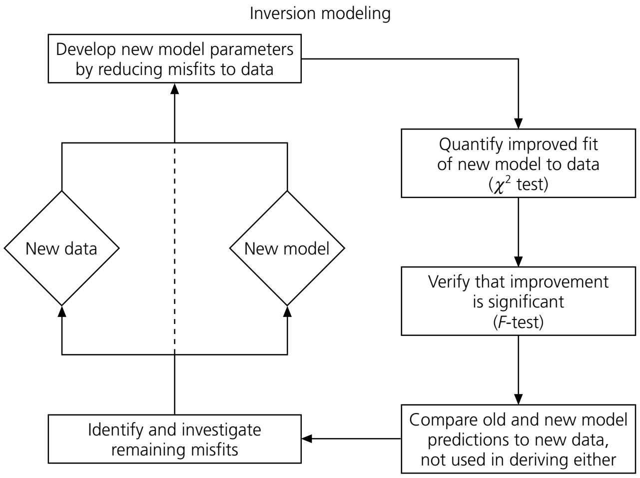


Fig. 1.1-8 Schematic illustration of how models of earth processes advance with time due to additional data and improved model parameterizations.

Over the years this process leads to a better understanding of how the earth works (Fig. 1.1-8). For example, Fig. 1.1-9 summarizes the development of global plate motion models, discussed in Chapter 5, that give the motion of the dozen or so major plates. The models are derived by inverting data consisting of the directions of plate motions along transform faults, the directions of plate motions during earthquakes, and the rates of plate motions shown by sea floor magnetic anomalies. Since 1972, when the first such model was made, the amount of available data has increased, and the data have become better, due to advances in seismology, sea floor imaging, and marine magnetic measurements. Similarly, the fit to the data has improved (or the misfit reduced) due both to the higher data quality and to improvements in the model, such as treating India and Australia as separate plates. Similar patterns of increased data and improved fit occur for many applications, including seismic velocity structure in the earth.

Many of the same issues surface when considering the models used to describe earth processes. For example, we will see that there are various models for what occurs at the core–mantle boundary or what causes earthquakes within downgoing plates at subduction zones. Such models assume that a particular set of physical processes occur, and show that for apparently plausible values of the (often unknown) relevant physical parameters, some behavior like that observed might be expected. Although these simple models attempt to reflect key aspects of the complex natural system, we often have no way of telling if and how well they succeed. Typically, various plausible models are suggested, all of which may in part be true and offer interesting insights into what may be occurring. The data often do not allow discrimination between them, so the model one prefers depends on one's geological instincts and prejudices, and models go in and out of vogue. A common scenario is for a model to become the consensus of the small group of researchers most interested in a problem, and then be challenged by fresh ideas or data from the outside. Hence, critically examining conventional wisdom often leads to discarding or modifying it, and so making progress in keeping with the

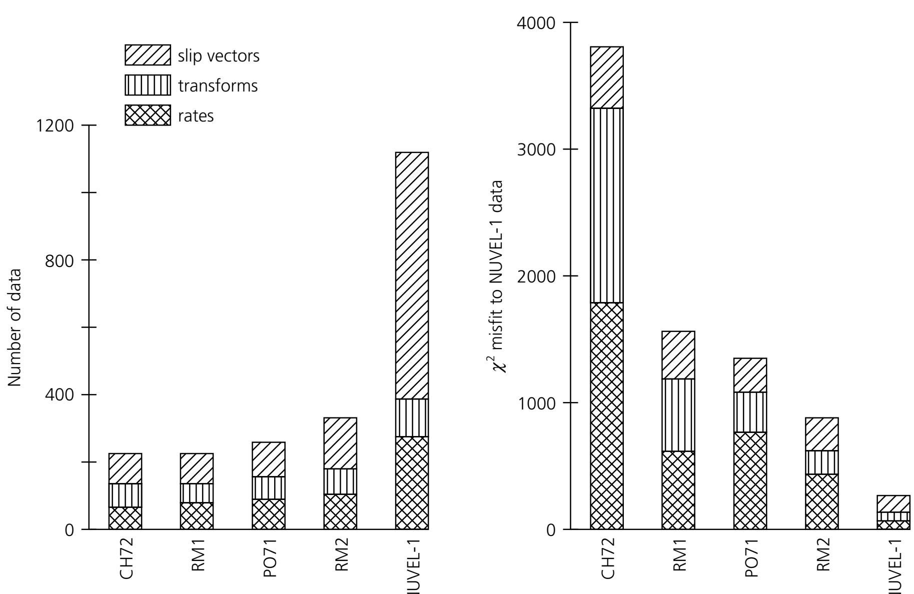


Fig. 1.1-9 Evolution of successive global plate motion models, as the amount of data increases and the misfit is reduced. *Left*: Number of data used to derive the models. Three types of data are inverted: earthquake slip vector azimuths, transform fault azimuths, and spreading rates. *Right*: The misfit to NUVEL-1 data for the various models. The vertical bars showing total misfit are separated into segments giving the misfit to each type of data. (DeMets *et al.*, 1990. *Geophys. J. Int.*, *101*, 425–78.)

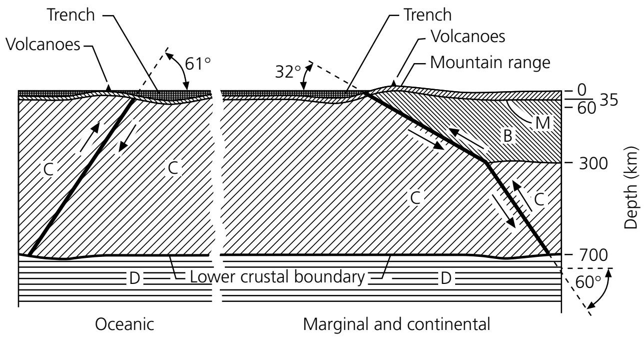


Fig. 1.1-10 Tectonic cartoon for oceanic and continental margin trenches, prior to the acceptance of plate tectonics. The association of dip-slip earthquakes with trenches, volcanism, and mountain ranges was recognized. Note the exaggeration of surface relief. (Benioff, 1955. From *Crust of the Earth*, ed. A. Poldervaart. Reproduced with permission of the publisher, the Geological Society of America, Boulder, CO. Copyright © 1955 Geological Society of America.)

ancient Jewish sages' observation that "the rivalry of scholars increases wisdom."[7] This process requires a constant cycle of learning and unlearning in which old models are discarded, even by those who helped create them, in favor of new models.

The classic geological example of advancing beyond conventional thinking is the plate tectonic revolution of the late 1960s. Although the idea of continental drift had been around for a long time and was strongly advocated by Alfred Wegener in 1915, it was not accepted by most of the geological community in the USA and Europe,[8] in part because seismological pioneer Harold Jeffreys argued that it was impossible. As a result, although it was recognized in the 1950s that earthquakes occurred on mid-ocean ridges that were young volcanic features and at deep sea trenches in association with volcanoes and mountain ranges (Fig. 1.1-10), their underlying nature was not understood. However, once paleomagnetic and marine geophysical data led to the recognition that oceanic lithosphere formed at mid-ocean ridges and subducted at trenches, the seismological observations made sense.

Thus, as in other sciences, progress in understanding seismological problems is typically incremental during "normal science" periods, in which we make small steady advances. Occasionally, however, exciting "paradigm shifts" occur when important new ideas change our views from our previous conventional thinking and permit great advances. This concept, developed by philosopher of science Thomas Kuhn (1962) for science-wide conceptual revolutions like the theory of plate tectonics, also describes progress in subfields. It is particularly apt in seismology, because many major faults move at most slightly for many years — and then break dramatically in large earthquakes.

[7] Alternative formulations of this idea include David Jackson's observation, (Fischman, 1992); "as soon as I hear 'everybody knows' I start asking 'does everybody know this, and how do they know it?'" the quotation used as the epigraph to this book by Nobel Laureate Peter Medewar; and the adage attributed to 1960s political activist Abbie Hoffman that "sacred cows make the best hamburger."

[8] Interestingly, many geologists in Southern Hemisphere countries like Australia and South Africa accepted continental drift early on and never abandoned it.

1.2 Seismology and society

Seismology impacts society through applications including seismic exploration for resources, earthquake studies, and nuclear arms control. These topics involve both scientific and public policy issues beyond our focus on using seismic waves to study earth structure, earthquakes, and plate tectonics. However, given the natural interest of these societal applications, we briefly discuss some issues in earthquake hazard analysis and nuclear test monitoring, in part to motivate our discussions of the basic science.

These topics have the interesting feature that the state of seismological knowledge influences policy, so scientific uncertainties have broad implications. The choice of earthquake preparedness strategies depends in part on how well earthquake hazards can be assessed, and nations' willingness to negotiate test ban treaties depend in part on their confidence that compliance can be verified seismologically. Seismology thus faces the challenge, familiar in other applications like global warming or biotechnology, of explaining both knowledge and its limits. Failure to do so can have embarrassing consequences. For example, since the 1960s the Japanese government has spent more than $1 billion on an earthquake prediction program premised on the idea that large earthquakes will be preceded by observable precursory phenomena, despite the fact that (as discussed shortly) many seismologists increasingly doubt that such phenomena exist. This approach has so far failed to predict destructive earthquakes, like that which struck the Kobe area in 1995, and has focused most of its efforts on areas other than those where these earthquakes occurred. Critics have thus argued that the program is scientifically weak, diverts resources that could be more usefully employed for basic seismology and earthquake engineering, and gives the public the misleading impression that earthquakes can currently be predicted. Based on the program's record to date, the government would have been wiser to listen to these critics and to have been more candid with the public.[1]

[1] Such issues were eloquently summarized by Richard Feynman's (1988) admonition after the loss of the space shuttle *Challenger*: "NASA owes it to the citizens from whom it asks support to be frank, honest, and informative, so these citizens can make the wisest decisions for the use of their limited resources. For a successful technology, reality must take precedence over public relations, because nature cannot be fooled."

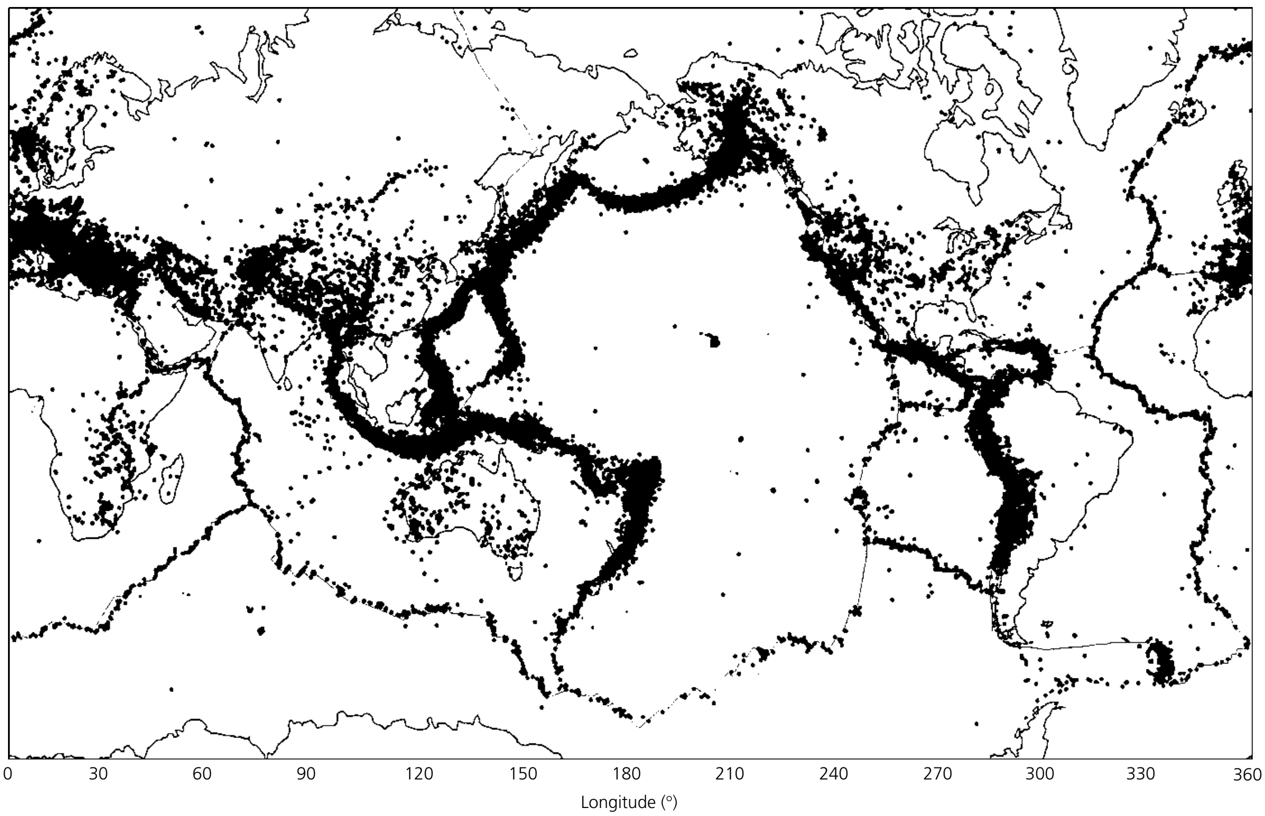


Fig. 1.2-1 Map showing epicenters of all earthquakes during 1963–95 with magnitudes of $m_b \geq 4$. Most earthquakes occur along the boundaries between tectonic plates. Where these boundaries are distinct, the earthquakes occur within narrow bounds. More diffuse plate boundaries, like the Himalayan plateau between India and China, show a much broader distribution of epicenters.

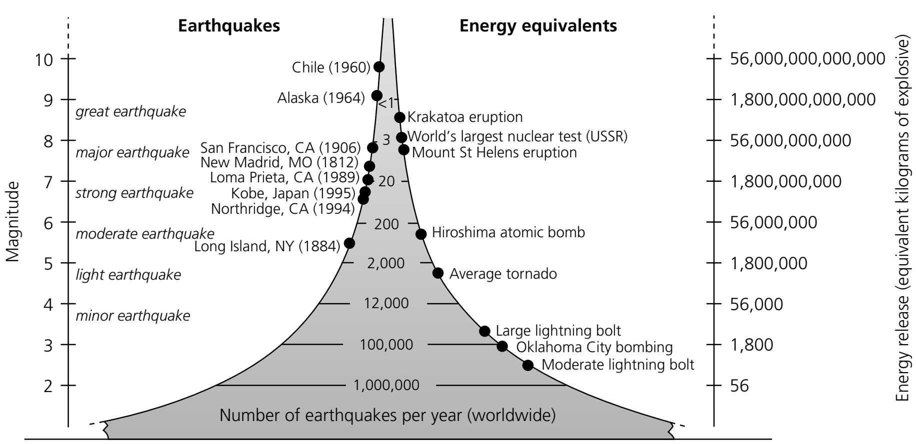


Fig. 1.2-2 Comparison of frequency, magnitude, and energy release of earthquakes and other phenomena. The magnitude used is moment magnitude, M_w. (After Incorporated Research Institutions for Seismology.)

1.2.1 Seismic hazards and risks

One of the primary motivations for studying earthquakes and seismology is the destruction caused by large earthquakes. In many parts of the world, seismic risks are significant, whether they are popularly recognized (as in Japan, where schools conduct earthquake drills) or not. Much of the challenge in assessing and addressing seismic hazards is that in any given area large earthquakes are relatively rare on human time scales, but can cause great destruction when they occur.

Earthquakes primarily occur at the boundaries where the 100 km-thick tectonic plates converge, diverge, or slide past each other. Although the plates move steadily, their boundaries are often "locked," and do not move most of the time. However, on time scales of a few hundred years, the boundary slips suddenly, and the accumulated motion is released in an earthquake. Figure 1.2-1 shows the locations of $m_b \geq 4$ earthquakes between 1963 and 1995. The earthquakes nicely define the plate boundaries, although some earthquakes also occur in *intraplate* regions, away from plate boundaries.

The energy released by large earthquakes is striking (Fig. 1.2-2). For example, the 1906 San Francisco earthquake involved about 4 m of slip on a 450 km-long fault, releasing about 3×10^{16} Joules[2] of elastic energy. This energy is equivalent to a 7 megaton nuclear explosion, much larger than the 0.012 megaton bomb dropped on Hiroshima. The largest recorded earthquake, the 1960 Chilean event in which about 21 m of slip occurred on a fault 800 km long and 200 km across, released about 10^{19} J of elastic energy, more than a 2000 Mt bomb. This earthquake released more energy than all the nuclear bombs ever exploded, the largest of which was 58 Mt. For comparison, the total global human annual energy consumption is about 3×10^{20} J.

Fortunately, the largest earthquakes are infrequent, because the energy released accumulates slowly over a long time. The San Francisco earthquake occurred on the San Andreas fault in northern California, part of the boundary along which the Pacific plate moves northward relative to the North American plate. Studies using the Global Positioning System satellites show that away from the plate boundary the two plates move by each other at a speed of about 45 mm/yr. Most parts of the San Andreas fault are "locked" most of the time, but slip several meters in a large earthquake every few hundred years. A simple calculation suggests that such earthquakes should occur on average about every 4000 mm/(45 mm/yr) or 90 years. The real interval is not uniform, for reasons that are unclear, and is longer, because some of the motion occurs on other faults.

Because plate boundaries extend for more than 150,000 km, and some earthquakes occur in plate interiors, earthquakes occur frequently somewhere on earth. As shown in Table 1.2-1, an earthquake of magnitude 7 occurs approximately monthly, and an earthquake of magnitude 6 or greater occurs on average every three days.[3] Earthquakes of a given magnitude occur about ten times less frequently than those one magnitude smaller. Because the magnitude is proportional to the logarithm of the energy released, most of the energy released seismically is in the largest earthquakes. A magnitude 8.5 event releases more energy than all the other earthquakes in a given year combined. Hence the hazard from earthquakes is due primarily to large (typically magnitude greater than 6.5) earthquakes.

Table 1.2-1 Numbers of earthquakes per year.

Earthquake magnitude (M_s)	Number per year	Energy released (10^{15} J/yr)
≥8.0	0–1	0–1,000
7–7.9	12	100
6–6.9	110	30
5–5.9	1,400	5
4–4.9	13,500	1
3–3.9	>100,000	0.2

Based upon data from the US Geological Survey National Earthquake Information Center. Energy estimates are based upon an empirical formula of Gutenberg and Richter (Gutenberg, 1959), and the magnitude scaling relations of Geller (1976), and are very approximate.

In assessing the potential danger posed by earthquakes or other natural disasters, it is useful to distinguish between *hazards* and *risks*. The hazard is the intrinsic natural occurrence of earthquakes and the resulting ground motion and other effects. The risk is the danger the hazard poses to life and property. Hence, although the hazard is an unavoidable geological fact, the risk is affected by human actions. Areas of high hazard can have low risk because few people live there, and areas of modest hazard can have high risk due to large populations and poor construction. Earthquake risks can be reduced by human actions, whereas hazards cannot (hence the US government's National Earthquake Hazards Reduction Program is, strictly speaking, misnamed).

These ideas are illustrated by Table 1.2-2, which lists some significant earthquakes and their societal consequences. As shown, some very large earthquakes caused no fatalities because of their remote location or deep focal depth. In general, the most destructive earthquakes occur where large populations live near plate boundaries. The highest property losses occur in developed nations where more property is at risk, whereas fatalities are highest in developing nations. Although the statistics are often imprecise, the impact of major earthquakes can be enormous. Estimates are that the 1990 Northern Iran shock killed 40,000 people, and that the 1988 Spitak

[2] The SI unit of energy is 1 Joule (J) = 1 Newton meter (N-m) = 10^7 ergs = 10^7 dyn-cm. Nuclear explosions are often described in megatons (Mt), equivalent to 1,000,000 tons of TNT or 4.2×10^{15} J.

[3] As part of his incorrect prediction of a magnitude 7 earthquake in the Midwest in 1990, I. Browning claimed that he had successfully predicted the 1989 Loma Prieta earthquake. In fact, he had said that near the date in question there would be an earthquake somewhere in the world with magnitude 6, a prediction virtually guaranteed to be true.

Table 1.2-2 Some notable and destructive earthquakes. (Values in this table are compiled from various sources, and different estimates have been reported, especially for older earthquakes.)

Location and date	Strength	Effects
Kourion, Cyprus July 21, 365	**X** MMI	Total destruction of this Greco-Roman city. Very large tsunami in the Mediterranean.
Basel, Switzerland October 18, 1356	**XI** MMI	Eighty castles destroyed over a wide area. 300 killed. Toppled cooking hearths caused fires that burned for many days.
Shansi, China January 23, 1556	**8** M_s (est.)	Collapse of cave dwellings carved into bluffs of soft glacial loess. 830,000 reported killed (worst ever). Near the 1920 Kansu earthquake (see below).
Port Royal, Jamaica June 7, 1692	**8** M_s (est.)	Widespread liquefaction caused one-third of Port Royal to spread and sink 4 m beneath the ocean surface. 2500 killed.
Lisbon, Portugal November 1, 1755	**≥8** M_s (est.)	Large tsunamis seen all around the Atlantic. Felt over 1,600,000 km^2. Algiers destroyed. 70,000 killed. Largest documented earthquake in Europe (though several Italian quakes have killed >150,000 in past 500 years).
New Madrid, MO Dec. 1811 to Feb. 1812	**7–7.4** M_s (est.)	Three large quakes (Dec. 16, 1811, Jan. 23, 1812, Feb. 7, 1812). Vertical movements up to 7 m. Widespread liquefaction. Changed course of Mississippi River. Felt over 5,000,000 km^2.
Charleston, SC August 31, 1886	**7.2** M_s (est.)	No previous seismicity observed in this area between 1680 and 1886. Felt over 5,000,000 km^2. 14,000 chimneys damaged or destroyed. 90% of buildings damaged/destroyed. 60 killed.
Sanriku, Japan June 15, 1896	**8.5** M_s (est.)	Tsunamis 35 m high washed away 10,000 houses and killed 26,000 along the Sanriku coast of Honshu. A similar Sanriku quake on March 2, 1933, killed 3000 with a 25 m high tsunami.
Assam, India June 12, 1897	**8.7** M_s (est.)	One of the largest quakes ever felt. 1500 killed. Extremely violent ground shaking. Other Himalayan events on April 4, 1905 (20,000 killed), January 15, 1934 (10,000 killed), and August 15, 1950 (Ms = 8.6, 1526 killed).
San Francisco, CA April 18, 1906	**7.8** M_s	About 4 m of slip on a 450 km-long fault. 28,000 buildings destroyed, largely by fires that burned for 3 days. 2500–3000 killed by fires (worst in USA).
Kansu, China December 16, 1920	**8.5** M_s	180,000 killed, largely by downslope flow of liquefied soil over more than 1.5 km.
Tokyo, Japan September 1, 1923	**8.2** M_s	Occurred in Sagami Bay, 80 km south of Tokyo. 134 separate fires merged to become a giant firestorm. 12 m tsunami hit shores of Sagami Bay. 143,000 killed.
Aleutian Islands, Alaska April 1, 1946	**7.4** M_s	Large tsunami destroyed a power station and caused $25 million in damage in Hilo, Hawaii, where it rose to 7 m in height.
Lituya Bay, Alaska July 10, 1958	**7.0** M_s	Massive landslides that slid into a local bay created a 60 m-high wave that washed up mountain sides as far as 540 m.
Hebgen Lake, MT August 17, 1959	**7.5** M_s	Extensive landslides, including one that dammed a river and created a lake. Reactivated 160 Yellowstone geysers. Vertical displacement up to 6.5 m. 28 killed.
Chile May 21, 1960	**9.5** M_w	Largest quake ever recorded. Fault area: 800 by 200 km. Slip: 21 m. Triggered eruption of Puyehue volcano. Massive landslides in Andes. Giant tsunami. 2000–3000 killed.
Alaska March 27, 1964	**9.1** M_w	2nd largest quake ever recorded. Fault area: 500 by 300 km. Slip: 7 m. Large tsunamis, and widespread liquefaction. 200,000 km^2 of crustal surface deformed. 131 killed.
Peru May 31, 1970	**7.8** M_s	Quake offshore caused large landslides. 30,000 killed, largely by 100,000,000 m^3 of rock and ice flowing down Andes mountain sides.
San Fernando Valley, CA February 9, 1971	**6.6** M_s	Felt over more than 200,000 mi^2. 65 killed. 1000 injured. More than $500 million in direct losses.
Haicheng, China February 4, 1975	**7.4** M_s	Successful prediction said to have led to an evacuation on the morning of the quake that possibly saved 100,000s of lives. 300–1200 killed.
Kalapana, Hawaii November 29, 1975	**7.1** M_s	South flank of Kiluea volcano slid seaward. 14.6 m-high tsunami on Hawaiian shores. Largest Hawaiian earthquake since a 1868 quake that caused 22 m-high tsunamis and killed 148.
Tangshan, China July 27, 1976	**7.6** M_s	Of a city of 1 million, >250,000 killed and 50,000 injured. Exact numbers speculative: fatalities may have exceeded the 1556 earthquake. In contrast to the 1975 Haicheng quake, this had no precursory behaviors.
Mexico City, Mexico September 19, 1985	**7.9** M_s	Strong shaking lasted for 3 minutes due to sedimentary lake-fill oscillations. 10,000 killed. 30,000 injured. $3 billion in damage.
Spitak, Armenia December 7, 1988	**6.8** M_s	Surface faulting showed 1.5 m of slip along a 10 km fault. 25,000 killed. 19,000 injured. 500,000 homeless. $6.2 billion in damages.
Loma Prieta, CA October 17, 1989	**7.1** M_s	Slip along San Andreas segment south of San Francisco. 63 killed, most from the collapse of an elevated freeway in Oakland. About $6 billion in damages. Disrupted 5th game of World Series.
Caspian Sea, Iran June 20, 1990	**7.7** M_s	100,000 structures damaged or destroyed. 40,000 killed. 60,000 injured. 500,000 left homeless. Over 700 villages destroyed, and another 300 damaged.
Luzon, Philippines July 16, 1990	**7.8** M_s	Major rupture of Digdig fault, causing many landslides and major surface faulting. Extensive soil liquefaction. 1621 killed. 3000 injured.
Landers, CA June 28, 1992	**7.3** M_w	Up to 6 m of horizontal displacement and 2 m of vertical displacement along a 70 km fault segment. 1 killed. 400 injured.

Table 1.2-2 (*cont'd*).

Location and date	Strength	Effects
Flores Island, Indonesia December 12, 1992	**7.8** M_s	Tsunami heights reached 25 m. Extensive shoreline damage, where tsunami run-up was up to 300 m. 2200 killed. 30,000 buildings destroyed.
Northridge, CA January 17, 1994	**6.7** M_w	Rupture on a blind thrust fault beneath Los Angeles. Many rock slides, ground cracks, and soil liquefaction. 58 killed. 7000 injured. 20,000 homeless. About $20 billion in damages.
Northern Bolivia June 9, 1994	**8.2** M_s	Largest deep earthquake ever (depth was 637 km). Felt as far away as Canada.
Kobe, Japan January 16, 1995	**6.8** M_s	5502 killed. 36,896 injured. 310,000 homeless. Massive destruction to world's 3rd largest seaport: 193,000 buildings, $100 billion in damages (highest to date).
NW of Balleny Islands March 25, 1998	**8.2** M_w	Largest oceanic intraplate earthquake ever. Occurred west of Australia–Pacific–Antarctic plate triple junction in a region that was previously aseismic.
Izmit, Turkey August 17, 1999	**7.4** M_s	5 m slip. 120 km rupture. 30,000 killed. $20 billion in economic loss. 12 major ($M > 6.7$) events this century have broken a total of 1000 km of the North Anatolian fault, including a 7.2 Mw aftershock on Nov. 12, 1999.
Chi-Chi, Taiwan September 21, 1999	**7.6** M_w	150 km south of Taipei. 2333 killed. 10,000 injured. >100,000 homeless. Extensive seismic monitoring in Taiwan makes this one of the best seismically sampled earthquakes. One of largest observed surface thrust scarps.

(Armenia) earthquake killed 25,000. Even in Japan, where modern construction practices are used to reduce earthquake damage, the 1995 Kobe earthquake caused more than 5000 deaths and $100 billion of damage. On average during the past century earthquakes have caused about 11,500 deaths per year. As a result, earthquakes have had a significant effect upon the history and culture of many regions.

The earthquake risk in the United States is much less than in many other countries because large earthquakes are relatively rare in most of the country and because of earthquake-resistant construction.[4] The most seismically active area is southern Alaska, a subduction zone subject to large earthquakes. However, the population there is relatively small, so the 1964 earthquake (the second largest ever recorded instrumentally) caused far fewer deaths than a comparable earthquake would have in Japan. The primary earthquake impact in recent years has been in California. The 1994 Northridge earthquake killed 58 people and caused about $20 billion worth of damage in the Los Angeles area, and the 1989 Loma Prieta earthquake that shook the San Francisco area during a 1989 World Series baseball game killed 63 people and did about $6 billion worth of damage. Both these earthquakes were smaller (magnitude 6.8 and 7.1, respectively) than the largest known to occur on the San Andreas fault, such as the 1906 San Francisco earthquake, which had a magnitude of about 7.8.

Compared to other risks, earthquakes are not a major cause of death or damage in the USA. Most earthquakes do little harm, and even those felt in populated areas are commonly more of a nuisance than a catastrophe. Since 1811, US earthquakes have claimed an average of nine lives per year (Table 1.2-3), putting earthquakes at the level of in-line skating

[4] Many seismologists have faced situations like explaining to apprehensive telephone callers that the danger of earthquakes is small enough that the callers' upcoming family vacations to Disneyland are not suicidal ventures.

Table 1.2-3 Some causes of death in the United States, 1996.

Cause of death	Number of deaths
Heart attack	733,834
Cancer	544,278
Stroke	160,431
Lung disease	106,143
Pneumonia/influenza	82,579
Diabetes	61,559
Motor vehicle accidents	43,300
AIDS	32,655
Suicide	30,862
Liver disease/cirrhosis	25,135
Kidney disease	24,391
Alzheimer's	21,166
Homicide	20,738
Falling	14,100
Poison	10,400
Drowning	3,900
Fires	3,200
Suffocation	3,000
Bicycle accidents	695
Severe weather[1]	514
In-line skating[2]	25
Football[2]	18
Skateboards[2]	10
Earthquakes (1811–1983),[3] per year	9
Earthquakes (1984–98), per year	9

[1] From the National Weather Service (property loss due to severe weather is $10–15 billion/yr, comparable to the Northridge earthquake, and that from individual hurricanes can go up to $25 billion).
[2] From the Consumer Product Safety Commission.
[3] From Gere and Shah (1984).
All others from the National Safety Council and National Center for Health Statistics.

or football,[5] but far less than bicycles, for risk of loss of life. Similarly, the $20 billion worth of damage from the Northridge earthquake, though enormous, is about 10% of the annual loss due to automobile accidents. As a result, earthquakes pose an interesting challenge to society because they cause infrequent, but occasionally major, fatalities and damage. Society seems better able to accept risks that are more frequent but where individual events are less destructive.[6]

Similar issues surface when society must decide the costs, benefits, and appropriateness of various measures to reduce earthquake risks. Conceptually, the issues are essentially those faced in daily life. For example, a home security system costing $200 per year makes sense if one anticipates losing $1000 in property to a burglary about every five years ($200/year), but not if this loss is likely only once every 25 years ($40/year). However, the analysis is difficult, because the limited historical record of earthquakes makes it hard to assess their recurrence and potential damage.

Seismology is used in various ways to try to mitigate earthquake risks. Studies of past earthquakes are integrated with other geophysical data to forecast the location and size of future earthquakes. These estimates help engineers design earthquake-resistant structures, and help engineers and public authorities estimate and prepare for future damage by developing codes for earthquake-resistant construction. Seismology is also used by the insurance industry to develop rates for earthquake insurance, which can reduce the financial losses due to earthquakes and provide the resources for economic recovery after a damaging earthquake. Rates can be based on factors including the nature of a structure, its location relative to active faults, and soil conditions. Homeowners and businesses then decide whether to purchase insurance, depending on their perceived risk and the fact that damages must exceed a deductible amount (10–15% of the insured value) before the insurance company pays. A complexity for the insurer is that, unlike automobile accidents, whose occurrence is relatively uniform, earthquakes or other natural disasters are rare but can produce concentrated damage so large as to imperil the insurer's ability to pay claims. Approaches to this problem include limits on how much a company will insure in a given area, the use of reinsurance by which one insurance company insures another, catastrophe bonds that spread the financial risk into the global capital market, and government insurance programs.

1.2.2 Engineering seismology and earthquake engineering

Most earthquake-related deaths result from the collapse of buildings, because people standing in an open field during a large earthquake would just be knocked down. Thus it is often stated that in general "earthquakes don't kill people; buildings kill people." As a result, proper construction is the primary method used to reduce earthquake risks. This issue is addressed by engineering seismology and earthquake engineering, disciplines at the interface between seismology and civil engineering. Their joint goal is to understand the earthquake ground motions that can damage buildings and other critical structures, and to design structures to survive them or at least ensure the safety of the inhabitants.

These studies focus on the strong ground motion near earthquakes that is large enough to do damage, rather than the much smaller and often imperceptible ground motions used in many other seismological applications. Two common measures are used to characterize the ground motion at a site. One is the *acceleration*, or the second time derivative of the ground motion. Accelerations are primarily responsible for building destruction. A house would be unharmed on a high-speed train going along a straight track, where there is no acceleration. However, during an earthquake the house will be shaken and could be damaged if the accelerations were large enough. These issues are investigated using seismometers called *accelerometers* that can operate during violent shaking close to an earthquake but are less sensitive to the smaller ground motion from distant earthquakes. The seismic hazard to a given area is often described by numerical models that estimate how likely an area is to experience a certain acceleration in a given time. For example, the hazard map in Fig. 1.2-3 predicts the maximum acceleration expected at a 2% probability in the next 50 years, or at least once during the next 2500 (50/0.02) years. These values are given as a fraction of "g," the acceleration of gravity (9.8 m/s^2).

A second way to characterize strong ground motion uses *intensity*, a descriptive measure of the effects of shaking. Table 1.2-4 shows values for the commonly used Modified Mercalli intensity (MMI) scale, which uses roman numerals ranging from I (generally unfelt) to XII (total destruction). Intensity is not uniquely related to acceleration, which is a numerical parameter that seismologists compute for an earthquake and engineers use to describe building effects. The table shows an approximate correspondence between intensity and acceleration, but this can vary. However, intensity has the advantage that it is inferred from human accounts, and so can be determined where no seismometer was present and for earthquakes that occurred before the modern seismometer was invented (about 1890). Although intensity values can be imprecise (a fallen chimney can raise the value for a large area), they are often the best information available about historic earthquakes. For example, intensity data provide much of what is known about the New Madrid earthquakes of 1811 and 1812 (Fig. 1.2-4). These large earthquakes are interesting in that they occurred in the relatively stable continental interior of the North American plate (Section 5.6). Historical accounts show that houses fell down (intensity X) in the tiny Mississippi river town of New Madrid, and several chimneys toppled (intensity VII) near St Louis. Intensities can be used to infer earthquake magnitudes, albeit with significant uncertainties. These data have been used to infer the magnitude (about 7.2 ±

[5] These figures are for American football; in other countries soccer, termed football there, is safer for players but more dangerous for spectators.

[6] For example, although considerable attention is paid to aviation disasters and safety, far more lives could be saved at far less cost by enforcing automobile seat belt laws.

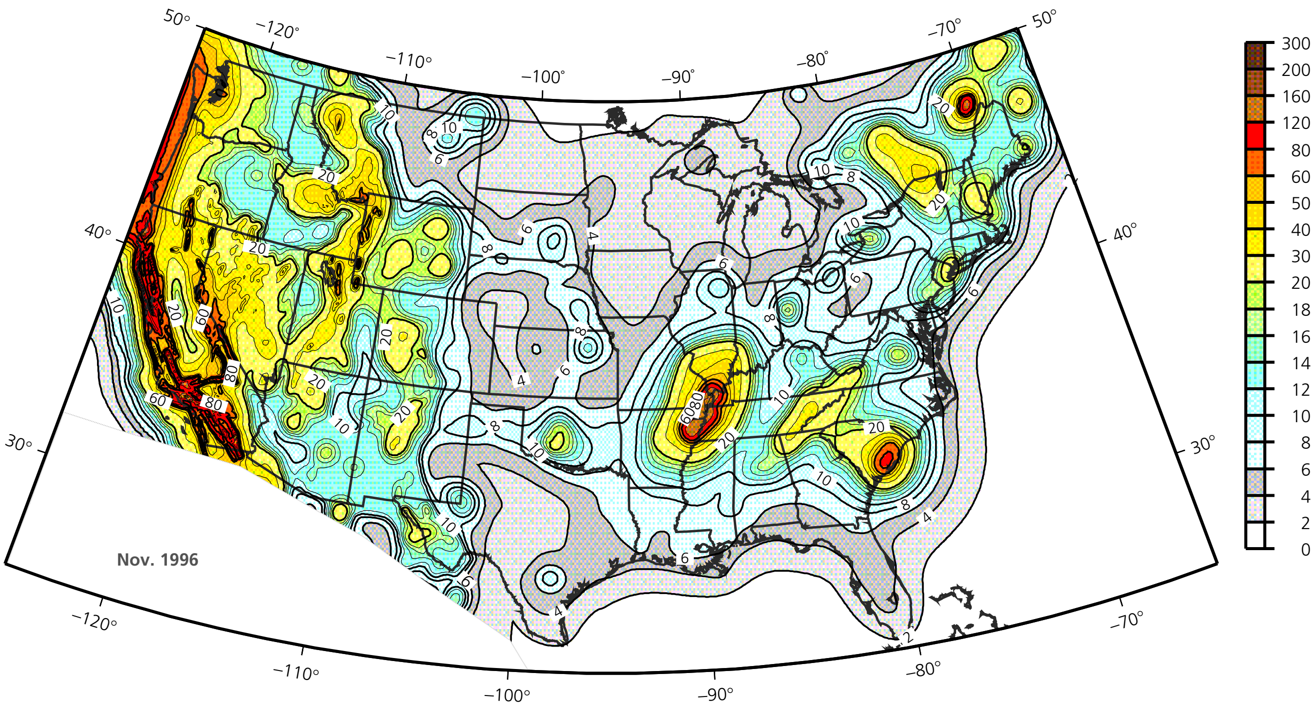


Fig. 1.2-3 A map of estimated earthquake hazards in the United States. The predicted hazards are plotted as the maximum acceleration of ground shaking expected at a 2% probability over a 50-year period. Although the only active plate boundaries are in the western USA, other areas are also shown as having significant hazards. (Courtesy of the US Geological Survey.)

0.3 in the study shown) and fault geometry of the historic earthquakes and to give insight into the effects of future ones.

The variation in ground motion with distance from an earthquake can be seen by plotting lines of constant intensity, known as *isoseismals*. Typically, as illustrated in Fig. 1.2-4, the intensity decays with distance from the earthquake. Similarly, strong motion data show that the variation in acceleration a with earthquake magnitude M and distance r from the earthquake can be described approximately by relations like

$$a(M, r) = b10^{cM} r^{-d}, \tag{1}$$

where b, c, and d are constants that depend on factors including the geology of the area in question, the earthquake depth and fault geometry, and the frequency of ground motion. Hence the predicted ground acceleration increases with earthquake magnitude and falls off rapidly with distance at a rate depending on the rock type. For example, rocks in the USA east of the Rocky Mountains transmit seismic energy better than those in the western USA (Section 3.7.10), so earthquakes in the East are felt over a larger area than earthquakes of the same size in the West (Fig. 1.2-5). Because the shaking decays rapidly with distance, nearby earthquakes can do more damage than larger ones further away.

The damage resulting from a given ground motion depends on the types of buildings. As shown in Fig. 1.2-6, reinforced concrete fares better during an earthquake than a timber frame, which does better than brick or masonry. Hence, as also shown in Table 1.2-4, serious damage occurs for about 10% of brick buildings starting above about intensity VII (about 0.2 g), whereas reinforced concrete buildings have similar damage only around intensity VIII–IX (about 0.3–0.5 g). Buildings designed with seismic safety features do even better. The worst earthquake fatalities, such as the approximately 25,000 deaths in the 1988 Spitak (Armenia) earthquake, occur where many of the buildings are vulnerable (Fig. 1.2-7). Hence a knowledgeable observer[7] estimated that an earthquake of this size would cause approximately 30 deaths in California. This estimate proved accurate for the 1989 Loma Prieta earthquake, which was slightly larger and killed 63 people.

Designing buildings to withstand earthquakes is a technical, economic, and societal challenge. Research is being directed to better understand how buildings respond to ground motion and how they should be built to best survive it. Because such design raises construction costs and thus diverts resources from other uses, some of which might save more lives at less cost or otherwise do more societal good, the issue is to assess the seismic hazard and choose a level of earthquake-resistant

[7] Ambraseys (1989).

Table 1.2-4 Modified Mercalli intensity scale.

Intersity	Effects
I	Shaking not felt, no damage: not felt except by a very few under especially favorable circumstances.
II	Shaking weak, no damage: felt only by a few persons at rest, especially on upper floors of buildings. Delicately suspended objects may swing.
III	Felt quite noticeably indoors, especially on upper floors of buildings, but many people do not recognize it as an earthquake. Standing automobiles may rock slightly. Vibration like passing of truck. Duration estimated.
IV	Shaking light, no damage: during the day felt indoors by many, outdoors by very few. At night some awakened. Dishes, windows, doors disturbed; walls make creaking sound. Sensation like heavy truck striking building. Standing automobiles rocked noticeably. (0.015–0.02 g)
V	Shaking moderate, very light damage: felt by nearly everyone, many awakened. Some dishes, windows, and so on broken; cracked plaster in a few places; unstable objects overturned. Disturbances of trees and poles, and other tall objects sometimes noticed. Pendulum clocks may stop. (0.03–0.04 g)
VI	Shaking strong, light damage: felt by all, many frightened and run outdoors. Some heavy furniture moved; a few instances of fallen plaster and damaged chimneys. Damage slight. (0.06–0.07 g)
VII	Shaking very strong, moderate damage: everybody runs outdoors. Damage negligible in buildings of good design and construction; slight to moderate in well-built ordinary structures; considerable in poorly built or badly designed structures; some chimneys broken. Noticed by persons driving cars. (0.10–0.15 g)
VIII	Shaking severe, moderate to heavy damage: damage slight in specially designed structures; considerable in ordinary substantial buildings with partial collapse; great in poorly built structures. Panel walls thrown out of frame structures. Fall of chimneys, factory stacks, columns, monuments, walls. Heavy furniture overturned. Sand and mud ejected in small amounts. Changes in well water. Persons driving cars disturbed. (0.25–0.30 g)
IX	Shaking violent, heavy damage: damage considerable in specially designed structures; well-designed frame structures thrown out of plumb; great in substantial buildings, with partial collapse. Buildings shifted off foundations. Ground cracked conspicuously. Underground pipes broken. (0.50–0.55 g)
X	Shaking extreme, very heavy damage: some well-built wooden structures destroyed; most masonry and frame structures destroyed with foundations; ground badly cracked. Rails bent. Landslides considerable from river banks and steep slopes. Shifted sand and mud. Water splashed, slopped over banks. (More than 0.60 g)
XI	Few, if any, (masonry) structures remain standing. Bridges destroyed. Broad fissures in ground. Underground pipelines completely out of service. Earth slumps and land slips in soft ground. Rails bent greatly.
XII	Damage total. Waves seen on ground surfaces. Lines of sight and level destroyed. Objects thrown into the air.

Note: Parentheses show the average peak acceleration in terms of g (9.8 m/s), taken from Bolt (1999).

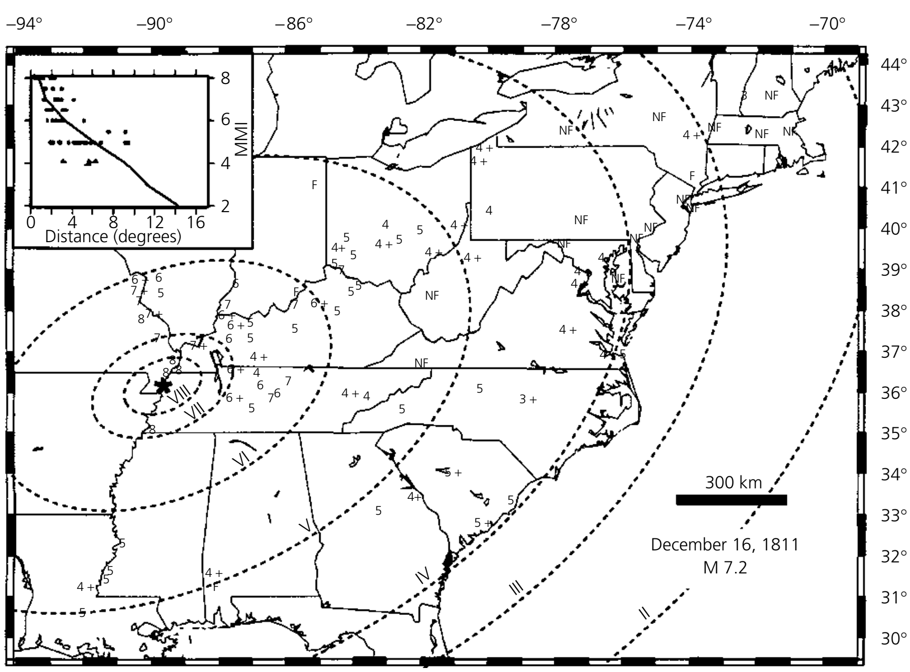


Fig. 1.2-4 Isoseismals for the first of the three largest earthquakes of the 1811–12 New Madrid earthquake sequence. Such plots, though based on sparse data, often provide the best assessment of historical earthquakes and of the effects of future ones. (After Hough *et al.*, 2000. *J. Geophys. Res.*, *105*, 23,839–64, Copyright by the American Geophysical Union.)

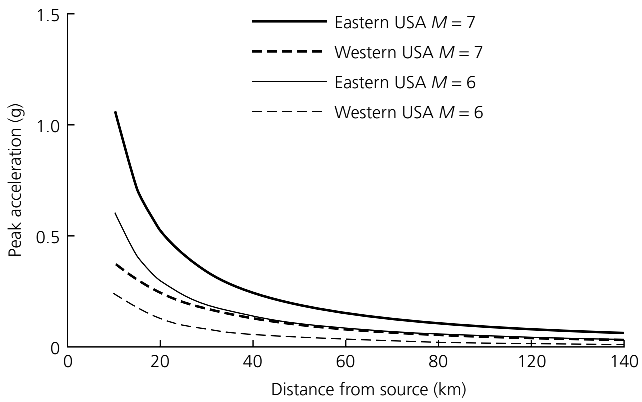


Fig. 1.2-5 Comparison of the predicted strong ground motion as a function of distance from magnitude 7 and 6 earthquakes in the eastern and western United States. Shaking from an earthquake in the east is comparable to that from one a magnitude unit larger in the west. The curves are computed from models by Atkinson and Boore (1995) and Sadigh *et al.* (1997).

Fig. 1.2-7 Five-story building in Spitak, Armenia, destroyed during the December 7, 1988, earthquake. The building was made from precast concrete frames that were inadequately connected. The failure of such buildings contributed greatly to the loss of 25,000 lives. (Courtesy of the US Geological Survey.)

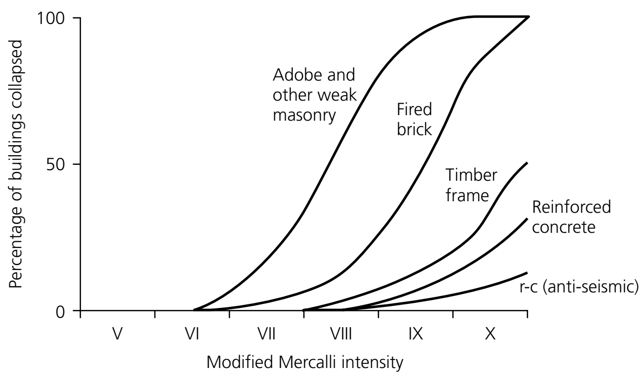


Fig. 1.2-6 Approximate percentage of buildings that collapse as a function of the intensity of earthquake-related shaking. The survival of buildings differs greatly for constructions of weak masonry, fired brick, timber, and reinforced concrete (with and without anti-seismic design). (After Coburn and Spence, *Earthquake Protection*, © 1992. Reproduced by permission of John Wiley & Sons Limited.)

construction that makes economic sense. Countries like the USA and Japan have the financial resources to study the effects of shaking on buildings, develop codes of appropriate building construction, and build structures to meet those codes. The task for building codes is to not be too weak, permitting unsafe construction and undue risks, or too strong, imposing unneeded costs and encouraging their evasion. Deciding where to draw this line is a complex policy issue for which there is no unique answer. Making the appropriate decisions is even more difficult in developing nations, many of which face serious hazards but have even larger alternative demands for resources that could be used for seismic safety. A classic example is the choice between building schools for towns without them or making existing schools earthquake-resistant.

A related issue is ensuring that buildings are built to the codes, given the tendency to evade expensive regulations designed to deal with events that are infrequent on a human time scale. For example, much damage occurred during large earthquakes in Turkey in 1999 because the building codes were not enforced. It has been reported that walls crumbled, revealing empty olive oil cans inserted during construction to save the costs of concrete.

Much of what has been learned about safe construction has been via trial and error. In California, the first major set of building codes was enacted following the 1933 Long Beach earthquake, which did $41 million worth of damage and killed 120 people. With successive destructive earthquakes, engineers have acquired a better sense of what works best, and building codes have been modified. For instance, buildings have become more resistant to the lateral shear that accompanies horizontal shaking with the use of shear walls consisting of concrete reinforced with steel. Similarly, measures have been developed to retrofit older buildings to increase their earthquake resistance.

An important factor for earthquake engineers is that structures resonate at different periods. Although the resonant period or periods depend on the specific building geometry and materials, they generally increase with an increase in the height or base width of a building. For example, typical houses or small buildings have periods of about 0.2 s, whereas a typical 10-story building has a period around 1 s. If the peak energy of ground motion is close to a building's resonant period, and the shaking continues long enough, the building may undergo large oscillations and be seriously damaged. This effect is like a swing — pushing at random intervals will likely stop the swing, whereas pushing repeatedly at its resonant period gives the person on it a good ride. Through this mechanism, an earthquake can destroy certain buildings and not others. Similarly, a building might collapse after a magnitude 7 earthquake, but remain standing after a magnitude 8 event with peak energy at a lower frequency. Sometimes damage occurs because adjacent buildings resonate out of phase, making their tops collide.

Another crucial factor for earthquake-resistant construction is the ground material of the site. Loose sediments and other weak rocks at the surface enhance ground motion compared to bedrock sites. As shown in Section 2.4.5, near-surface sediments can increase ground displacements by more than an order of magnitude. For instance, during the 1989 Loma Prieta earthquake, areas that sustained the worst damage corresponded to ones of high risk identified on the basis of subsurface geology. The failures of buildings in the Marina district, the Bay Bridge, and the Nimitz freeway all occurred on sedimentary layers.

An example of these effects occurred in 1985 in Mexico City, which is built on the sedimentary fill of an ancient lake that has dried up since the time of the Aztecs. A magnitude 7.9 earthquake at the subduction zone to the west caused the sedimentary basin to shake for more than 3 minutes (an unusually long time) at a dominant period of about 2 s. The worst damage was sustained by buildings with 6–15 stories, which had resonant periods of 1–3 s. Shorter or taller buildings were less damaged because they did not resonate with the ground shaking. This damage pattern has repeated for successive earthquakes.

1.2.3 Highways, bridges, dams, and pipelines

Buildings are not the only challenge for earthquake-resistant construction. Highways, bridges, parking structures, landfills, dams, pipelines, and power plants present additional problems. Many of these structures are crucial to society, so considerable effort is made to ensure that they will survive earthquakes.

Elevated highways often fail during earthquakes. Most of the lives lost during the 1989 Loma Prieta earthquake were due to the collapse of the Nimitz freeway in Oakland. In Los Angeles, the I-5 freeway was built to withstand a large earthquake, but parts were destroyed during the 1971 San Fernando earthquake. These were rebuilt, but parts collapsed again during the 1994 Northridge shock. A dramatic highway failure occurred during the 1995 Kobe earthquake, when a 20 km length of an expressway supported by large concrete piers fell over, crushing many cars and trucks.

Similar problems beset bridges, as illustrated in the 1989 Loma Prieta earthquake. The Bay Bridge connecting San Francisco and Oakland is a double-deck bridge built in 1936 with little flexibility and rests on sedimentary rocks. A large piece of the upper span collapsed during the earthquake (Fig. 1.2-8), and the bridge was closed for months for repairs. By contrast, the Golden Gate Bridge, a suspension bridge built into bedrock, was designed to withstand a large amount of shaking and fared well.

Fig. 1.2-8 Damage to the Bay Bridge, connecting San Francisco and Oakland, from the October 17, 1989, Loma Prieta earthquake. The bridge is of old construction (1936), and its supports rest in sedimentary fill that amplifies ground shaking. (Courtesy of the US Geological Survey.)

The failure of dams due to earthquakes poses considerable risk, as illustrated by the near-failure of the lower Van Norman dam during the 1971 San Fernando earthquake. A segment of the dam 600 m long broke and slid into the reservoir (Fig. 1.2-9), lowering the dam by 10 m and leaving it only 1.5 m above the water. Fortunately, the area had been suffering from a drought, and the reservoir was only half full. Eighty thousand people living below the dam were evacuated, and the reservoir was quickly drained. The dam was replaced by a more modern dam that suffered only minor cracking during the 1994 Northridge earthquake.

Dams have the special problem that they can cause earthquakes. This seems counter-intuitive, because the added weight of the water should increase the pressure on the rock below and inhibit faulting, because the two sides of the fault are pressed together harder, requiring a greater force to overcome the friction. However, it seems that the water impounded by dams sometimes flows into the rock, lowering the friction across faults and making rupture easier. The effect can be noticeable; seismicity associated with the man-made lake in Koyna, India, seems to follow a seasonal curve, being more active following the rainy season when reservoir levels are higher. One earthquake in 1967 was large enough to kill 200 people. The possibility of reservoir-induced earthquakes is thus considered when designing dams.

The greatest cause of earthquake-related death and destruction, other than the collapse of buildings, is fire. An important contributor to this problem is that water pipelines can rupture, making fire fighting harder. In the 1906 San Francisco earthquake, many buildings were damaged by the shaking, but fires that lasted three days are thought to have done ten times more

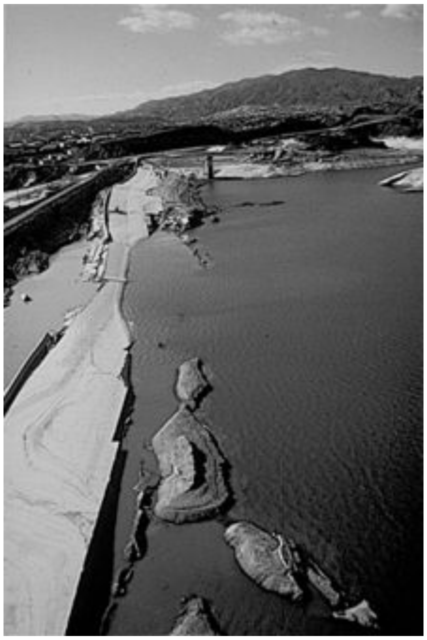


Fig. 1.2-9 Failure of the lower Van Norman dam that occurred during the February 9, 1971, San Fernando valley earthquake. Flooding did not occur because the region had been experiencing a drought, and the water level was low. (Courtesy of the US Geological Survey.)

damage (Fig. 1.2-10). Following the 1923 Tokyo earthquake, fires caused by overturned cooking stoves spread rapidly through the city and were unstoppable, due to ruptured water pipes. Many of the over 140,000 deaths resulted from fire, including a fire storm that engulfed 40,000 people who fled to an open area to escape collapsing buildings. In modern cities, natural gas pipelines can rupture, allowing flammable gas to escape and ignite. After the 1994 Northridge and 1995 Kobe earthquakes, both of which happened at night, the wide outbreaks of fires were the first way that rescue efforts could identify the areas that sustained the greatest damage. People in earthquake-prone areas are taught to turn off the gas supply to their homes if they smell gas after a large earthquake.

Fig. 1.2-11 Aerial view of Valdez, Alaska, showing the inundation of the coastline following the great 1964 earthquake. The resulting tsunami was as high as 32 m in places. (National Geophysical Data Center. Courtesy of the US Department of the Interior.)

1.2.4 Tsunamis, landslides, and soil liquefaction

Spectacular exceptions to the truism that "earthquakes don't kill people, buildings kill people" include tsunamis, landslides, avalanches, and soil liquefaction. Earthquake hazard planning thus includes identifying sites where these risks are present.

Tsunamis are large water waves that occur when portions of the sea floor are displaced by volcanic eruptions, submarine landslides, or underwater earthquakes (Fig. 1.2-11). Tsunamis are not noticeable as they cross the ocean, but can be amplified dramatically upon reaching the shore. The 1896 Sanriku (Japan) earthquake caused 35 m-high tsunamis that washed away 10,000 homes and killed 26,000 people. Hawaii is especially susceptible to tsunamis from earthquakes around the Pacific rim. Tsunamis from the 1960 Chilean earthquake killed 61 people in Hawaii, and the 1946 Alaska earthquake created a 7 m-high tsunami that washed over and short-circuited a power station, plunging Hilo into darkness. To address these risks, tsunami warning systems have been developed that assess

Fig. 1.2-10 Fires burning in San Francisco five hours after the April 18, 1906, earthquake. Many buildings received little damage from the earthquake, but were destroyed by the fires that burned out of control for three days. (Courtesy of the National Geophysical Data Center.)

Fig. 1.2-12 Landslide along California State Highway 17 in the Santa Cruz mountains, caused by shaking from the 1989 Loma Prieta earthquake. The landslide blocked the major commuter route between Santa Cruz and San Jose. (Courtesy of the US Geological Survey.)

Fig. 1.2-13 Damage to apartment buildings caused by soil liquefaction during the June 16, 1964, Niigata (Japan) earthquake. About a third of the city sank by as much 2 m as a result of sand compaction. (Courtesy of the National Geophysical Data Center.)

the likelihood that a large earthquake will generate a tsunami and issue warnings before the tsunami reaches distant areas.

Ground shaking in areas with steep topography can cause destructive landslides and avalanches (Fig. 1.2-12). For example, a 1970 earthquake in Peru caused rock and ice landslides that traveled downhill at speeds of 300 km/hr, burying villages and killing 30,000 people.

Another earthquake hazard involves *liquefaction*, a process by which loose water-saturated sands behave like liquids when vigorously shaken. Under normal conditions, the sand grains are in contact with each other, and water fills the pore spaces between them. Strong shaking moves the grains apart, so the soil behaves like a fluid slurry similar to "quicksand." Buildings can sink, otherwise undamaged, during the few seconds of peak ground shaking, and end up permanently stuck when the shaking stops and the soil resolidifies. A classic example is the tilting and sinking of buildings in Niigata, Japan, during a 1964 earthquake (Fig. 1.2-13).

Ground consisting of loose wet sediment is most susceptible to liquefaction. Sometimes the sand is ejected out of the surface as *sand blows*. This happened in the Marina district of the San Francisco waterfront during the 1989 Loma Prieta earthquake. Ironically, some of the material that erupted from the ground was building rubble from the 1906 San Francisco earthquake that had been bulldozed into the bay to make new waterfront property.

Liquefaction can be widespread and devastating, involving large downslope movements of soil called *lateral spreading*. In the 1920 Kansu, China, earthquake, downslope flows traveled over 1.5 km, killing 180,000 people. During the 1964 Alaska earthquake, parts of the Turnagain Heights section of Anchorage liquefied and collapsed. A dramatic example occurred on the island of Jamaica due to a magnitude 8 earthquake in 1692, where much of the town of Port Royal, built upon sand, sank about 4 m beneath the ocean. For years afterward, people on boats in the harbor could see houses below.

1.2.5 Earthquake forecasting

Reducing earthquake risks via resistant construction relies on identifying regions prone to earthquakes and estimating, even if crudely, how likely earthquakes are to occur and what shaking they might produce. Thus earthquake forecasting involves both scientific issues and the related question of how society can best use what seismology can provide.

Before addressing the predictions of earthquakes, it is useful to consider predictions for other geophysical processes. For example, severe storms are predicted in several ways. The first are long-term average forecasts: Chicagoans expect winter snowstorms, whereas Miamians expect fall hurricanes. Public authorities, power companies, homeowners, and businesses use the historical record of storms to prepare for them. Although surprises occur, long-term forecasting is generally adequate to ensure that needed resources (snow plows, salt) are available, whereas funds are not wasted on unneeded preparations (snow plows in Miami). Second, short-term weather forecasting often can identify conditions under which a storm is likely to form soon. Third, once formed, storms are tracked in *real time*, so people are often warned a day or more in advance to make preparations.

Similarly, volcanic hazard assessment begins with the location of volcanoes that are active or have been so recently (in geological terms). Based on the eruption history taken from historical accounts and the geologic record, long-term forecasts can be made. Short-term predictions are made using various phenomena that precede major eruptions: rising magma causes ground deformation, small earthquakes, and the release of volcanic gases. Finally, small eruptions usually precede a large one, making it possible to issue real-time warnings. Hence the record of volcanic predictions, though not perfect,[8] is reasonably good. The area around Mt St Helens was evacuated before the giant eruption of May 18, 1980, reducing the loss of life to only 60 people, including a geologist studying the volcano and citizens who refused to leave. The largest eruption of the second half of the twentieth century, Mt Pinatubo in the Philippines, destroyed over 100,000 houses and a nearby US Air Force base, yet only 281 people died because of evacuations during the preceding days.

Seismologists would like to do as well for earthquakes. We would like to be able to forecast where they are on average likely to occur in years to come, predict them a few years to hours before they occur, and issue real-time warnings after an earthquake has occurred in situations where such a warning would be useful. However, the record of seismology in these areas is mixed. To date there has been some success in long-term forecasting, little if any in short-term prediction, and some in real-time warning.

Earthquake forecasting, discussed in Section 4.7.3, estimates the probability that an earthquake of a certain magnitude will occur in a particular area during a specific time. For instance, a forecast might be a 25% probability of a magnitude 7 or greater earthquake occurring along the San Francisco segment of the San Andreas fault in the next 30 years. Forecasting uses the history of earthquakes on the fault and other geophysical information, such as the crustal motions measured using the Global Positioning System, to predict its likely future behavior. While forecasting is not relevant to short-term earthquake preparations, it is important in the enactment of building codes for earthquake-resistant construction, which are costly and require justification. Such forecasting is already successful in general ways; knowing that the San Andreas and nearby faults will be the sites of recurrent earthquakes has prompted building codes that are a major reason why the 1989 Loma Prieta and 1994 Northridge earthquakes caused few casualties.

Going beyond general forecasts is more difficult. For example, the probabilistic hazard map for the USA in Fig. 1.2-3 predicts a general pattern of higher hazards in areas of known past large earthquakes. Most of these, in California and Nevada, the Pacific Northwest, and Utah, are in the western USA, in the broad boundary zone between the Pacific and North American plates. In addition, high hazards are predicated in parts of the interior of the continent, near Charleston, South Carolina, and the New Madrid seismic zone in the Midwest. The map attempts to quantify this risk in terms of the maximum expected acceleration (recall that 0.2 g corresponds approximately to the onset of significant building damage) during a time interval. Such maps are made by assuming where and how often earthquakes will occur, how large they will be, and then using ground motion models like those in Fig. 1.2-5 to predict how much ground motion they will produce. Because these factors are not well understood, especially in intraplate regions where large earthquakes are rare, hazard estimates have considerable uncertainties.[9] For example, the high hazard predicted for parts of the Midwest, exceeding that in San Francisco or Los Angeles, results from specific assumptions, and alternative assumptions yield quite different estimates (Fig. 1.2-14).

Similarly, hazard estimates depend on the probability and hence recurrence time considered. Where the largest earthquakes are expected about every 200 years — for example, near a plate boundary as in California — a hazard map predicting the maximum acceleration expected at a 10% probability in the next 50 years, or at least once during the next 500 (50/0.1) years, will be similar to one for 2% probability in the next 50 years, or at least once during the next 2500 (50/0.02) years, because each portion of plate boundary is expected to rupture at least once in 500 years. However, the two maps would differ significantly where large earthquakes are less frequent — for example, in an intraplate region like the New Madrid zone (Sections 4.7.1, 5.6.3). This issue is important in choosing building codes because typical buildings have a useful life of about 50 years.

Because earthquakes are infrequent on a human time scale, it will be a long time before we know how well such estimates, which combine long-term earthquake forecasts and ground motion predictions, actually describe future earthquakes. Nonetheless, such estimates are used for purposes such as developing building codes and setting insurance rates. As a result, how to make meaningful predictions and hazard estimates, communicate their uncertainties to the public, and best use them for policy is a topic of discussion relevant not just to seismology but to the other earth sciences as well.

A key scientific challenge for hazard estimation is that the process determining when large earthquakes recur is unclear. The underlying basis for seismic forecasting is the principle of *elastic rebound* (Section 4.1). In this model, large-scale crustal motions, in most cases due to plate motions, slowly build up stress and strain across locked faults. When the stress reaches a critical threshold, seismic slip occurs along the fault, and the stress immediately drops. The process then begins again. The repeat time for these earthquakes depends on the rate at which crustal motions load the fault and the properties of the rocks that control when it slips.

[8] In 1982, uplift of the volcanic dome and other activity near the resort town of Mammoth Lakes, California, suggested that an eruption might be imminent. Geologists issued a volcano alert, resulting in significant tensions with local business leaders. When no eruption occurred, geologists were the target of much local anger, and the county supervisor who arranged for an escape route in the event of a volcanic eruption was recalled in a special election.

[9] Earthquake risk assessment has been described as "a game of chance of which we still don't know all the rules" (Lomnitz, 1989).

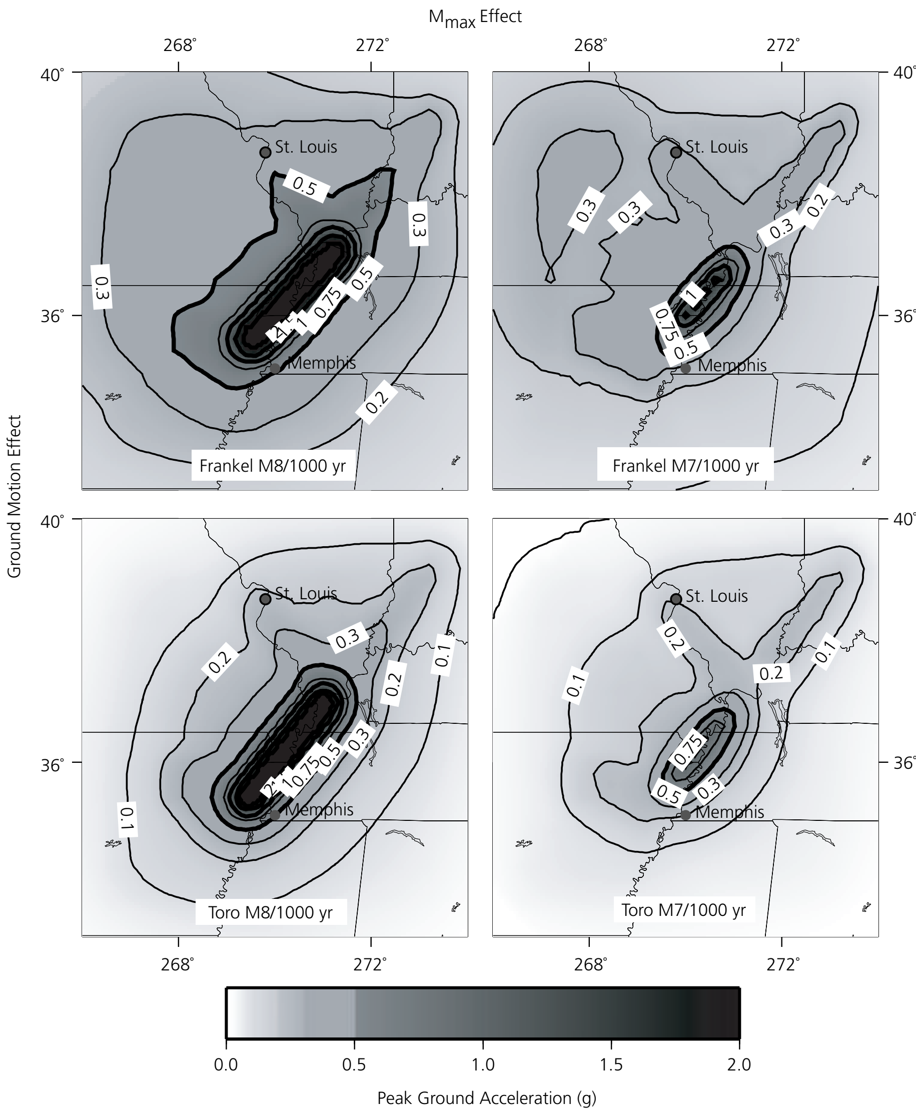


Fig. 1.2-14 Comparison of the predicted seismic hazard (peak ground acceleration expected at 2% probability in 50 years) from New Madrid seismic zone earthquakes for alternative parameter choices. Rows show the effect of varying the magnitude of the largest expected New Madrid fault earthquakes from 8 to 7, which primarily affects the predicted acceleration near the fault. Columns show the effect of two different ground motion models ("Frankel" and "Toro") which affect the predicted acceleration over a larger area. (Newman *et al.*, 2001. © Seismological Society of America. All rights reserved.)

This idea implies that the history of large past earthquakes in an area should indicate the probable time of the next one. Naturally, the longer the history available, the better. Unfortunately, the duration of earthquake cycles is typically long compared to the approximately 100-year history of instrumental seismology. In some parts of the world, like China and Japan, historical records extend well into the past, whereas in the USA, the historic record is shorter. The earthquake history can be extended by *paleoseismology*, a branch of geology that studies the past history of faults. One of the best examples is the use of geological data to infer the history of large earthquakes on a major southern segment of the San Andreas fault. The last major earthquake recorded at a site at Pallett Creek, California, the 1857 Fort Tejon earthquake, is known from historical records to have caused shaking with an intensity of XI. The faulting is recorded by disruptions of sedimentary strata, including sand blows where material erupted during the earthquake. Sand blows and other structures from previous earthquakes were dated with radiometric carbon-14 methods, giving the dates of previous earthquakes. Despite the many uncertainties involved with these methods, including uncertainties in radiometric dating and the effects of climate variations and burrowing animals, the data show that faulting has recurred over the past thousands of years. However, assessing the size of past earthquakes and whether some earthquakes were missed is difficult.

The results can be surprising. For instance, large earthquakes near Pallett Creek appear to have occurred approximately in the years 1857, 1812, 1480, 1346, 1100, 1048, 997, 797, 734, and 671. Because the average time between events is 132 years,

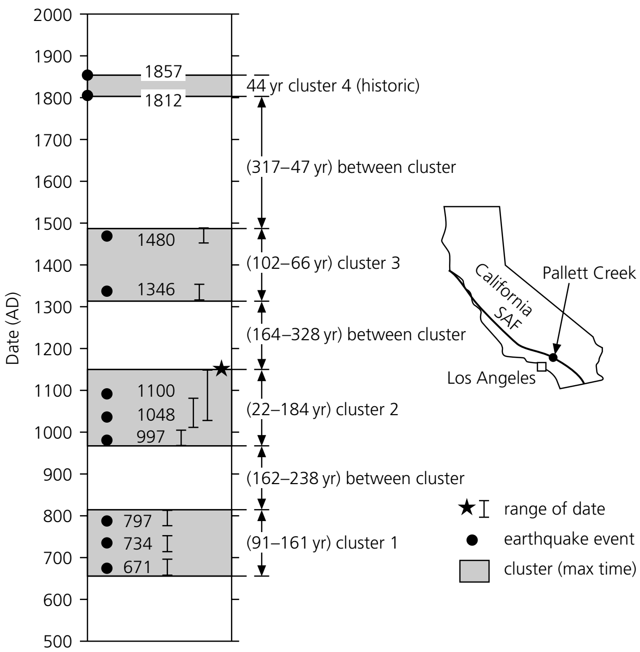


Fig. 1.2-15 Paleoseismic time series of earthquakes along the San Andreas fault near Pallett Creek, California, inferred from sedimentary deposits by Sieh *et al.* (1989). The sequence shows earthquake clusters separated by longer time intervals, illustrating the complexity of earthquake recurrence. (Keller and Pinter, *Active Tectonics: earthquakes, uplift, and the landscape*, © 1996. Reprinted by permission of Pearson Education.)

we might have expected the next large earthquake around the year 1989. However, the intervals between earthquakes vary from 45 years to 332 years, with a standard deviation of 105 years. Thus, given these data right after the 1857 earthquake, the simplest view would be that the earthquake would likely recur between 1885 and 2093. However, the time history suggests that something more complicated is going on (Fig. 1.2-15), as illustrated by the fact that the standard deviation of the recurrence time is similar to its mean. It looks as if the earthquakes are clustered: three earthquakes between 671 and 797, then a 200-year gap, then three between 997 and 1100, followed by a 246-year gap. Hence, using the earthquake history to forecast the next big earthquake is challenging, and the study's authors concluded in 1989 that one could estimate the probability of a similar earthquake before 2019 as only somewhere in the range 7–51%. For example, if the cluster that included the 1812 and 1857 earthquakes is over, then it may be a long time until the next big earthquake there.

The variability of recurrence times is striking because these data span for a long time history (10 earthquake cycles) on a plate boundary where the plate motion causing the earthquake is steady. The history of most faults is known only for the past few cycles, and the Pallett Creek data imply that these may not be representative of the long-term pattern. The recurrence may be even more complicated for earthquake zones within plates, many of which seem to act for only a few earthquake cycles, and others of which may be one-time events. Research, some of which is discussed in Section 5.7, is going on to investigate this complexity.

Even with the dates of previous major earthquakes, it is difficult to predict when the next will occur, as illustrated by the segment of the San Andreas fault near Parkfield, California. Compared to the southern segment just discussed, or the northern segment on which the 1906 earthquake occurred, the Parkfield segment is characterized by smaller earthquakes that occur more frequently and appear much more periodic. Earthquakes of magnitude 5–6 occurred in 1857, 1881, 1901, 1922, 1934, and 1966. The average recurrence interval is 22 years, and a linear fit to these dates made 1988 the likely date of the next event. In 1985, it was predicted at the 95% confidence level that the next Parkfield earthquake would occur before 1993, which was the USA's first official earthquake prediction. A comprehensive observing system was set up to monitor electrical resistivity, magnetic field strength, seismic wave velocity, microseismicity, ground tilting, water well levels and chemistry (especially radon content), and motion across the fault. The well-publicized experiment[10] hoped to observe precursory behavior, which seemed likely because surface cracks were observed 10 days before the 1966 earthquake and a pipeline ruptured 9 hours before the shock, and to obtain detailed records of the earthquake at short distances. As of 2002, the earthquake had not yet happened, making the current interval (35 years and growing) the longest yet observed between earthquakes there. The next Parkfield earthquake will eventually occur, but its non-arrival to date illustrates both the limitations of the statistical approaches used in the prediction (including the omission of the 1934 earthquake on the grounds that it was premature and should have occurred in 1944) and the fact that even in the best of circumstances nature is not necessarily cooperative or easily predicted. For that matter, it is unclear whether the Parkfield segment of the San Andreas fault shows such unusual quasi-periodicity because it differs from other parts of the San Andreas fault (in which case predicting earthquakes there might not be that helpful for other parts), or whether it results simply from the fact that, given enough time and different fault segments, essentially random seismicity can yield apparent periodicity somewhere. As is usual with such questions, only time will tell.

Such seismic forecasting involves the concept of *seismic gaps*, discussed further in Sections 4.7.3 and 5.4.3. The idea is that a long plate boundary like the San Andreas or an oceanic trench ruptures in segments. We would thus expect steady plate motion to cause earthquakes that fill in gaps and occur at relatively regular intervals. However, the Pallett Creek and

[10] The costs involved (more than $30 million) led *The Economist* magazine (Aug. 1, 1987) to argue that "Parkfield is geophysics' Waterloo. If the earthquake comes without warnings of any kind, earthquakes are unpredictable and science is defeated. There will be no excuses left, for never has an ambush been more carefully laid."

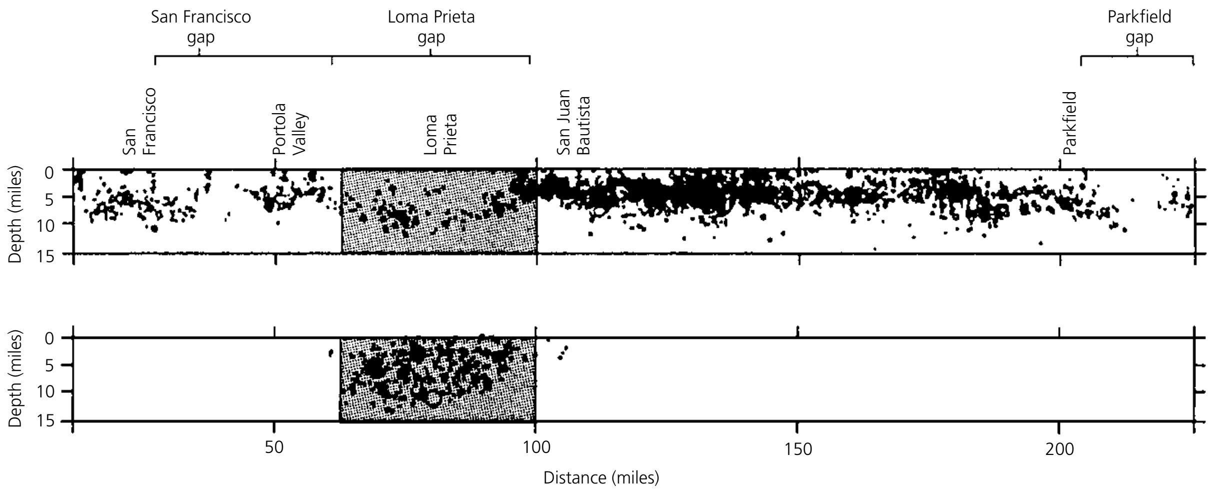


Fig. 1.2-16 Cross-section of the seismicity along the San Andreas fault before (top) and after (bottom) the 1989 Loma Prieta earthquake. This earthquake, whose rupture began at the large circle in the lower figure and is marked by the aftershocks (small circles), has been interpreted as filling a seismic gap along the San Andreas fault, although other interpretations have also been made. (Courtesy of the US Geological Survey.)

Parkfield examples show that the earth is more complicated. Some earthquakes may fit the gap idea; the 1989 Loma Prieta earthquake and its aftershocks have been interpreted as filling a gap along the San Andreas fault (Fig. 1.2-16), although the fact that the earthquake differed from the expected fault geometry has also been interpreted as making it different from the expected gap-filling earthquake. In other areas, however, the gap hypothesis has not yet proved successful in identifying future earthquake locations significantly better than random guessing. Faults deemed likely to rupture have not done so, and earthquakes sometimes occur on faults that were either unknown or considered seismically inactive. Understanding if, where, and when the gap hypothesis is useful is thus an active research area. Until it is resolved, it is unclear whether it is better to assume that all segments of a given fault are equally likely to rupture, making the probability of a major earthquake independent of time, or whether the segment that ruptured longest ago should have since accumulated the greatest elastic strain, and therefore be most likely to rupture next. This issue is important for hazard estimates.

In summary, several factors make earthquake forecasting difficult. In the meteorological case, storms occur frequently on human time scales, and we believe that we understand their basic physics. By contrast, the cycle of earthquakes on a given fault segment is long on a human time scale. Thus there are only a few places with a time history long enough to formulate useful hypotheses (recall that even the Pallett Creek 1000-year history shows major complexity). Moreover, because forecasts must be tested by their ability to predict future earthquakes, a long time will be needed to convincingly test models of earthquake recurrence and hazards. Even worse, the fundamental physics of earthquake faulting is not yet understood. Clearly, the process is complex. Earthquakes are at best only crudely periodic, and sometimes appear instead to cluster in time. Faults display a continuum of behavior from locking, to slow aseismic creep, to earthquakes. Thus the theoretical and experimental study of rock deformation and its application to earthquake faulting is an active field of research (Section 5.7).

1.2.6 *Earthquake prediction*

Earthquake prediction is defined as specifying within certain ranges the location, time, and size of an earthquake a few years to days before it occurs. Prediction is an even more difficult problem than long-term forecasting. A common analogy is that although a bending stick will eventually snap, it is hard to predict exactly when. To do so requires either a theoretical basis for knowing when the stick will break, given a history of the applied force, or observing some change in physical properties that immediately precedes the stick's failure.

Because little is known about the fundamental physics of faulting, many attempts to predict earthquakes have searched for *precursors*, observable behavior that precedes earthquakes. To date, as discussed next, this search has proved generally unsuccessful. As a result, it is unclear whether earthquake prediction is even possible. In one hypothesis, all earthquakes start off as tiny earthquakes, which happen frequently, but only a few cascade via a random failure process into large earthquakes.[11]

[11] This hypothesis draws on ideas from nonlinear dynamics or chaos theory, in which small perturbations can grow to have unpredictable large consequences. These ideas were posed in terms of the possibility that the flap of a butterfly's wings in Brazil might set off a tornado in Texas, or in general that minuscule disturbances do not affect the overall frequency of storms but can modify when they occur (Lorenz, 1993).

In this view, because there is nothing special about those tiny earthquakes that happen to grow into large ones, the interval between large earthquakes is highly variable, and no observable precursors should occur before them. If so, earthquake prediction is either impossible or nearly so.

Support for this view comes from the failure to observe a compelling pattern of precursory behavior before earthquakes. Various possible precursors have been suggested, and some may have been real in certain cases, but none have yet proved to be a general feature preceding all earthquakes, or to stand out convincingly from the normal range of the earth's variable behavior. Although it is tempting to note a precursory pattern after an earthquake based on a small set of data and to suggest that the earthquake might have been predicted, rigorous tests with large sets of data are needed to tell whether a possible precursory behavior is real and correlates with earthquakes more frequently than expected purely by chance. Most crucially, any such pattern needs to be tested by predicting future earthquakes.

One class of precursors involves *foreshocks*, earthquakes that occur before a main shock. Many earthquakes, in hindsight, have followed periods of anomalous seismicity. In some cases, there is a flurry of *microseismicity*: very small earthquakes like the cracking that precedes a bent stick's snapping. In other cases, there is no preceding seismicity. However, faults often show periods of either elevated or nonexistent microseismicity that are not followed by a large earthquake. Alternatively, the level of microseismicity before a large event can be unremarkable, occurring at a normal low level. The lack of a pattern highlights the problem with possible earthquake precursors: to date, no changes that might be associated with an upcoming earthquake are consistently distinguishable from the normal variations in seismicity that are not followed by a large earthquake.

Another class of possible precursors involves changes in the properties of rock within a fault zone preceding a large earthquake. It has been suggested that as a region experiences a buildup of elastic stress and strain, microcracks may form and fill with water, lowering the strength of the rock and eventually leading to an earthquake. This effect has been advocated based on data showing changes in the level of radon gas, presumably reflecting the development of microcracks that allow radon to escape. For example, the radon detected in groundwater rose steadily in the months before the 1995 Kobe earthquake, increased further two week before the earthquake, and then returned to a background level (Fig. 1.2-17).

A variety of similar observations have been reported. In some cases, the ratio of *P*- and *S*-wave speeds in the region of an earthquake has been reported to have decreased by as much as 10% before an earthquake. Such observations would be consistent with laboratory experiments, and would reflect cracks opening in the rock (lowering wave speeds) due to increasing stress and later filling (increasing wave speeds). However, this phenomenon has not been substantiated as a general phenomenon. Similar difficulties beset reports of a decrease in the electrical resistivity of the ground before some earthquakes, consistent with large-scale microcracking. Changes in the amount and composition of groundwater have also been observed. For example, a geyser in Calistoga, California, changed its period between eruptions before the 1989 Loma Prieta and 1975 Oroville, California, earthquakes.

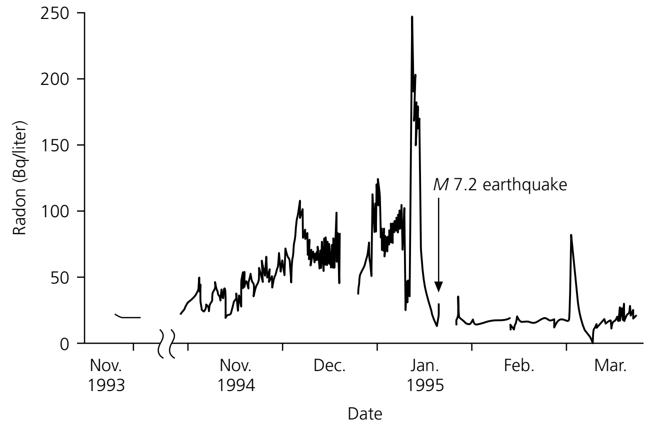


Fig. 1.2-17 Radon within groundwater before and after the January 16, 1995, Kobe earthquake in Japan. (Igarashi *et al.*, 1995. Reprinted with permission from *Science*, *269*, 60–1. Copyright 1995, American Association for the Advancement of Science.)

Efforts have also been made to identify ground deformation immediately preceding earthquakes. The most famous of these studies was the report in 1975 of 30–45 cm of uplift along the San Andreas fault near Palmdale, California. This highly publicized "Palmdale Bulge" was interpreted as evidence of an impending large earthquake and was a factor in the US government's decision to launch the National Earthquake Hazards Reduction Program aimed at studying and predicting earthquakes. However, the earthquake did not occur, and reanalysis of the data implied that the bulge had been an artifact of errors involved in referring the vertical motions to sea level via a traverse across the San Gabriel mountains. Subsequent studies, using newer and more accurate techniques including the Global Positioning System satellites, satellite radar interferometry, and borehole strainmeters have not yet convincingly detected precursory ground deformation.

An often-reported precursor that is even harder to quantify is anomalous animal behavior. What the animals are sensing (high-frequency noise, electromagnetic fields, gas emissions) is unclear. Moreover, because it is hard to distinguish "anomalous" behaviors from the usual range of animal behaviors, most such observations have been "postdictions," coming after rather than before an earthquake.

Despite these difficulties, Chinese scientists are attempting to predict earthquakes using precursors. Chinese sources report a successful prediction in which the city of Haicheng was evacuated in 1975, prior to a magnitude 7.4 earthquake that damaged more than 90% of the houses. The prediction is said to have been based on precursors, including ground deformation, changes in the electromagnetic field and groundwater levels, anomalous animal behavior, and significant foreshocks. However, in the following year, the Tangshan earthquake occurred not too far away without precursors. In minutes, 250,000 people died, and another 500,000 people were injured. In the following month, an earthquake warning in the Kwangtung province caused people to sleep in tents for two months, but no earthquake occurred. Because foreign scientists have not yet been able to assess the Chinese data and the record of predictions, including both false positives (predictions without earthquakes) and false negatives (earthquakes without predictions), it is difficult to evaluate the program.

In summary, despite tantalizing suggestions, at present there is still an absence of reliable precursors. The frustrations of this search have led to the wry observation that "it is difficult to predict earthquakes, especially before they happen." Most researchers thus feel that although earthquake prediction would be seismology's greatest triumph, it is either far away or will never happen. However, because success would be of enormous societal benefit, the search for methods of earthquake prediction will likely continue.

1.2.7 Real-time warnings

Some recent efforts are directed to the tractable goal of real-time warnings, where seismometers trigger an immediate warning if a set of criteria is met. For tsunamis, the warning may be several hours in advance, which is enough time for preparations. This is because tsunamis travel more slowly than seismic waves. A P wave travels from Alaska to Hawaii in about 7 minutes, whereas a tsunami traveling at about 800 km/hr across the ocean takes 5.5 hours. After the damage done to Hilo by the 1946 Alaska earthquake, the Seismic Sea Wave Warning System was organized for countries that rim the Pacific Ocean. Information from seismometers and tide gauges was phoned to the Tsunami Warning Center in Honolulu, Hawaii, which issued tsunami alerts if necessary.[12] Tsunami warning systems have since become more automated, using real-time digital seismic data to locate large earthquakes and derive information about their magnitudes, depths, and focal mechanisms. An assessment can be made of the likelihood of a tsunami, which usually results from vertical motion at the sea floor.

The situation is much more complicated with seismic waves. Although local seismic networks can automatically and immediately locate an earthquake and assess if it is hazardous, the warning time is short. For example, a warning after a major earthquake on the New Madrid fault system instantly relayed via Internet or radio to St Louis would arrive about 40 seconds before the first seismic waves. Seismologists, engineers, and public authorities are thus discussing what might be done with such short warning times. Although such times would not permit evacuations, certain steps might be useful. For example, real-time warnings are used in Japan to stop high-speed trains, and it may be practical to have gas line shut-off valves or other automatic responses connected to such a system. The questions are whether the improved safety justifies the cost and whether the risk of false alarms is serious.

A related approach is to provide authorities with near-real-time information, including data on the distribution of shaking, immediately after major earthquakes. Seismic networks are working to provide emergency management services with information that can help direct the needed response to the most affected areas during the chaotic few hours after a large earthquake, when the location and extent of damage are often still unclear.

[12] Serious or older television viewers may recall the episode of *Hawaii 5-0* in which criminals force the center to issue a spurious tsunami warning to prompt evacuation of downtown Honolulu and facilitate a robbery.

1.2.8 Nuclear monitoring and treaty verification

Another important societal application of seismology is the monitoring of nuclear testing. Although atomic physics destabilized world politics through the invention of the atomic bomb, seismology has partially restabilized it. Throughout the cold war between the USA and the Soviet Union, seismology helped verify that treaties were being observed.

The role of seismology in nuclear monitoring began in 1957 when the USA detonated RAINIER, the first underground nuclear explosion. By the early 1960s it became clear that radioactive elements produced by atmospheric nuclear testing posed significant health threats. In 1963, 116 nations signed the Limited Test Ban Treaty, which banned nuclear testing in the atmosphere, in the oceans, and in space, and required testing to occur underground. At about this time, the US Air Force helped fund the deployment of the World Wide Standardized Seismographic Network (WWSSN). WWSSN stations provided important information for monitoring nuclear testing and a wealth of data that played a major role in modern geophysical seismology.

In 1976, countries began to abide by the Threshold Test Ban Treaty, which limited the size of underground nuclear tests to 150 kt (equivalent to 150 kilotons of TNT). Before then, the largest atmospheric test had been 58 Mt, and the largest underground test had been 4.4 Mt. Figure 1.2-18 shows the yields estimated seismologically for underground nuclear tests carried out by the Soviet Union. Although it was initially thought that some of the post-1976 explosions were greater than 150 kt, this turned out to reflect the different geologies of the western USA and central Asia. The conversion of seismic body wave magnitude m_b values into TNT yields was calibrated using the Nevada test site, but the western US crust is more seismically attenuating than the more stable Soviet sites in Kazakhstan and Novaya Zemlya (see Section 3.7.10). The yields of explosions in kilotons, Y, can be related to the observed seismic magnitudes by

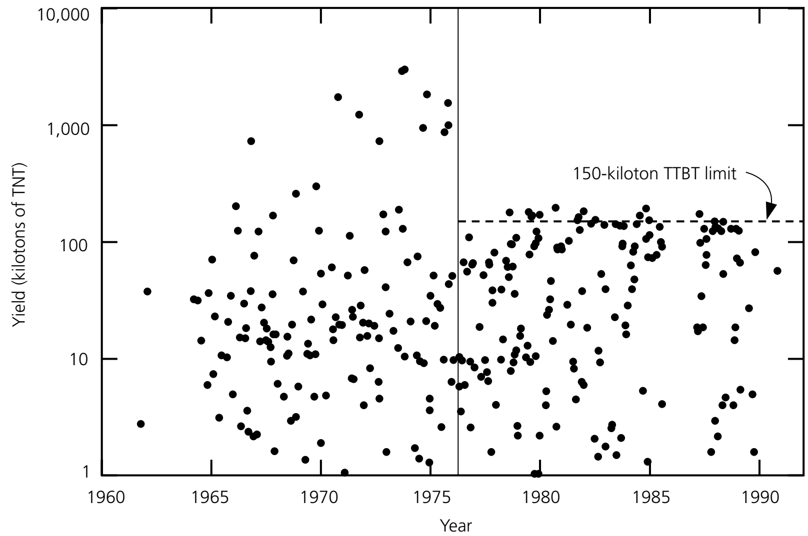


Fig. 1.2-18 Yields of underground nuclear tests carried out by the Soviet Union, determined through seismically observed m_b magnitudes. After the Threshold Test Ban Treaty (TTBT), seismology verified that the Soviet Union was in general compliance with the 150-kiloton limit. Data courtesy of P. Richards (personal communication).

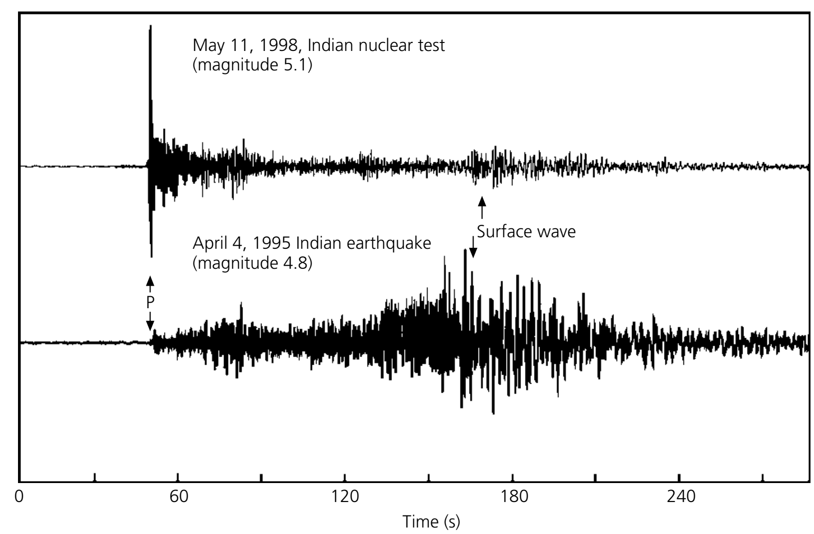


Fig. 1.2-19 Seismograms showing the differences between an earthquake and an explosion. For shallow earthquakes, in this case an m_b 4.8 shock in India, the P wave is much smaller than the surface waves. By contrast, the initial P wave is the largest arrival for explosions like this Indian nuclear test. Data recorded at Nilore, Pakistan. (Courtesy of the Incorporated Research Institutions for Seismology.)

$$m_b = C + 0.75 \log Y, \quad (2)$$

but the constant differs for Nevada ($C = 3.95$) and Kazakhstan ($C = 4.45$). With these corrections, it appears that the Soviet Union complied with the treaty.

Monitoring nuclear tests requires distinguishing them from earthquakes. Examples of the differences are shown in Fig. 1.2-19 for an earthquake and an explosion in India. Earthquakes occur by slip across a fault, generating large amounts of shear wave energy and hence large surface waves. By contrast, explosions involve motions away from the source, and so produce far less shear wave energy. Hence, for bombs the surface waves are dwarfed by the initial P wave. This difference is the basis for discrimination between earthquakes and explosions. A plot of M_s vs m_b (Fig. 1.2-20) separates earthquakes, which generate more surface wave energy (M_s), from the explosions, which generate more body (P) wave energy (m_b).

The challenge of seismic monitoring has increased in recent years. Since 1996 the USA has abided by the Comprehensive Test Ban Treaty (CTBT), which bans all nuclear testing, preventing the development of new nuclear weapons. Thus the focus of US monitoring efforts has expanded to include smaller countries around the world.[13] There is also the need to identify possible smaller nuclear tests, including those by terrorists. Hence seismic monitoring must identify explosions less than 1 kt, which have a magnitude of 4–4.5 (Eqn 2). This requires locating and identifying more than 200,000 earthquakes and additional mining explosions every year.

[13] A strategy described as "In God we trust, all others we verify."

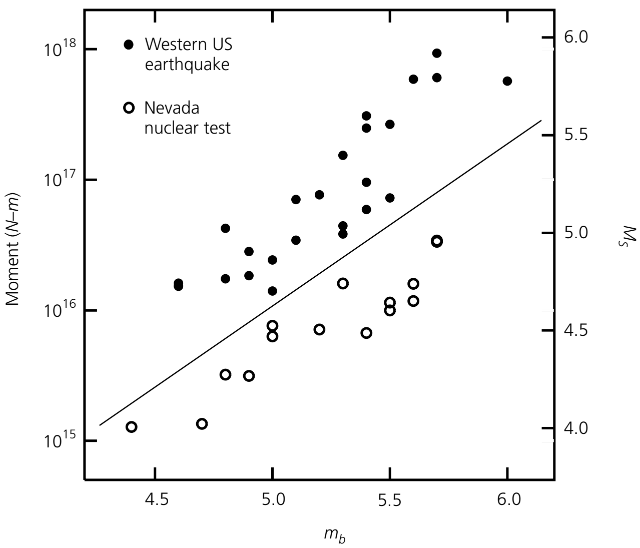


Fig. 1.2-20 Body wave magnitudes (m_b) versus surface wave magnitude (M_s) and seismic moment (M_0) for a set of earthquakes and explosions in the western USA. Because the *P* waves of explosions are very large, as shown in the previous figure, they have anomalously high m_b values for a given source energy (represented by M_0). A comparison of m_b and M_0 can thus discriminate between earthquakes and nuclear explosions. (After Al-eqabi *et al.*, 2001. © Seismological Society of America. All rights reserved.)

An important part of this effort is the International Monitoring System (IMS), whose aim is to detect, locate, and identify nuclear detonations that occur underground, underwater, or above ground. To do this, the IMS will combine seismological, hydroacoustic, and infrasound networks. Underwater nuclear tests create sound waves that travel efficiently through the ocean (Section 2.5.8), so a network of hydroacoustic stations will be established, with some sites using underwater hydrophones and others on islands to observe seismic phases that are generated when the oceanic acoustic waves reach land. Nuclear tests in the atmosphere will be detected by the infrasonic (frequencies less than 20 Hz, below the human hearing range) sound waves they generate. The IMS infrasound network will consist of small arrays of microphones that can determine the direction in which the infrasonic waves are traveling, so detection at multiple stations will identify the source of the waves.

Because most clandestine tests would likely occur underground, seismic stations will be a vital part of the IMS. The IMS seismic network will have 50 primary stations with three-component broadband seismometers. About half of these sites will be augmented with local arrays of short-period vertical-component sensors. Data will be telemetered in real time, so that there is no delay in monitoring. An auxiliary network of 120 broadband stations, distributed over 61 countries and largely based on existing networks, will aid in discrimination and replace malfunctioning primary stations.

Further reading

The seismological topics introduced in this chapter are discussed elsewhere in the text, so references are given in the appropriate sections. Many other references exist for the topics of societal interest discussed here.

Popular accounts of issues related to earthquakes include Gere and Shah (1984), Bolt (1999), and Brumbaugh (1999). Introductory treatments dealing with earthquakes and volcanoes from the point of view of the geology and hazards include Alexander (1993), Kovach (1995), and Sieh and LeVay (1998). The World Wide Web contains a wealth of general earthquake information; sites to start at include *http://www.scec.org*, *http://www.seismosoc.org*, *http://www.iris.edu*, and *http://earthquake.usgs.gov*. Specific issues related to volcano prediction studies at Mammoth Lakes are discussed by Sieh and LeVay (1998) and Hill (1998). For discussions of paleoseismology and geological effects of earthquakes, see Keller and Pinter (1996) and Yeats *et al.* (1997). The role of seismology in the plate tectonic revolution is discussed by Cox (1973) and Menard (1986); the general idea of scientific revolutions as “paradigm shifts” is given by Kuhn (1962).

Issues of assessing probabilities and uncertainties are discussed by Ekeland (1993); Henrion and Fischoff (1986) analyze the history of measurements of physical constants. Probabilistic seismic hazard analysis is discussed by Reiter (1990), Hanks and Cornell (1994), and Hanks (1997). The US Geological Survey National Seismic Hazard maps are described by Frankel *et al.* (1996), and a global hazard map is described by Shedlock *et al.* (2000). Uncertainties in earthquake probabilities for California are discussed by Savage (1991). Real-time seismology applications to earthquake risk mitigation are discussed by Kanamori *et al.* (1997). Sarewitz *et al.* (2000) discuss general issues of prediction and policy for the earth sciences, including earthquake prediction. Geschwind (2001) reviews the history of seismic risk mitigation and earthquake prediction policies in the USA.

A considerable volume of scientific literature addresses earthquake prediction, often arguing whether either a specific approach or any method can predict earthquakes. Turcotte (1991) gives a general review of many aspects of the topic, and Geller (1997) summarizes the history of earthquake prediction efforts, including that at Parkfield and the Palmdale Bulge. Geller *et al.* (1997) and Evans (1997) argue that earthquakes are unpredictable; Lomnitz (1994), Wyss *et al.* (1997), and Sykes *et al.* (1999) argue the other side. The Parkfield earthquake prediction experiment is summarized by Roeloffs and Langbein (1994); Davis *et al.* (1989) and Savage (1993) discuss the limitations of the statistical approach used. The controversy over the seismic gap hypothesis is discussed by Stein (1992); Kagan and Jackson (1991) and Jackson and Kagan (1993) argue against the hypothesis, and Nishenko and Sykes (1993) argue for it.

Earthquake engineering is discussed by Bray (1995), Chopra (1995), Krinitzsky *et al.* (1993), and Wiegel (1970). A good World Wide Web site to start at is *http://www.eeri.org*, which also provides an introduction to earthquake insurance. Issues in natural disaster insurance are discussed by Michaels *et al.* (1997).

Bolt (1976), Sykes and Davis (1987), Richards and Zavales (1990), and Lay (1992) discuss seismic verification of nuclear testing. More description of the Comprehensive Test Ban Treaty can be found at *http://pws.ctbto.org*.