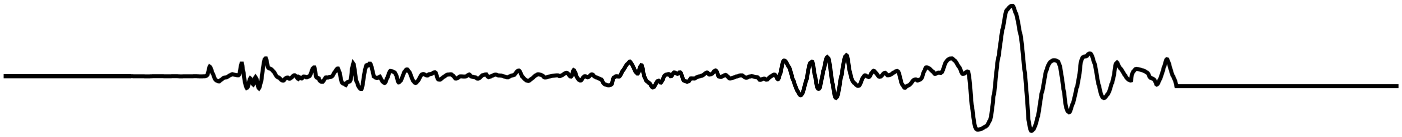


An Introduction to Seismology, Earthquakes, and Earth Structure

Seth Stein
Department of Geological Sciences, Northwestern University, Evanston, Illinois

Michael Wysession
Department of Earth and Planetary Sciences, Washington University, St Louis, Missouri

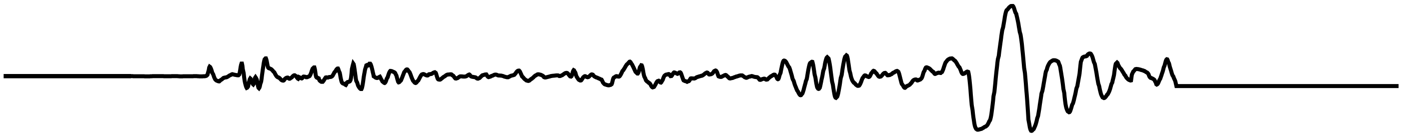
Blackwell
Publishing

AN INTRODUCTION TO SEISMOLOGY, EARTHQUAKES, AND EARTH STRUCTURE

To future generations of earth scientists — may their enthusiasm and creativity keep seismology vibrant and exciting

I cannot give any scientist of any age better advice than this: the intensity of the conviction that a hypothesis is true has no bearing on whether it is true or not. The importance of the strength of our conviction is only to provide a proportionally strong incentive to find out if the hypothesis will stand up to critical examination.

Sir Peter Medawar, *Advice to a Young Scientist*, 1979



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BLACKWELL PUBLISHING
350 Main Street, Malden, MA 02148-5020, USA
9600 Garsington Road, Oxford OX4 2DQ, UK
550 Swanston Street, Carlton, Victoria 3053, Australia

First published 2003 by Blackwell Publishing Ltd.

4 2005

Library of Congress Cataloging-in-Publication Data

Stein, Seth.
An introduction to seismology, earthquakes, and earth structure / Seth Stein, Michael Wysession.
p. cm.
Includes bibliographical references and index.
ISBN 0-86542-078-5 (pb. : alk. paper)
1. Seismology. 2. Geology, Structural. 3. Earthquakes. I. Wysession, Michael. II. Title.

QE534.3 .S74 2002
551.22—dc21
2001052639

ISBN-13: 978-0-86542-078-6 (pb. : alk. paper)

A catalogue record for this title is available from the British Library.

Set in 9½/11½ pt Sabon
by Graphicraft Ltd, Hong Kong
Printed and bound in India
by Replika Press Pvt Ltd, Kundli

The publisher's policy is to use permanent paper from mills that operate a sustainable forestry policy, and which has been manufactured from pulp processed using acid-free and elementary chlorine-free practices. Furthermore, the publisher ensures that the text paper and cover board used have met acceptable environmental accreditation standards.

For further information on
Blackwell Publishing, visit our website:
www.blackwellpublishing.com

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Preface

Science is only worth doing if it is interesting and fun. Hence the goal of a textbook is to interest students in a subject, convince them it is worth the effort required to learn about it, and help them do so. We have tried here to do all three.

For seismology, these should be easy. It is hard to imagine topics more interesting than the structure and evolution of a planet, as manifested by phenomena as dramatic as earthquakes. Our goal is to address them via an introduction to seismology, which is one of the cornerstones of the modern earth sciences. Seismology has been defined as the study of earthquakes and associated phenomena, or the study of elastic waves propagating in the earth. By integrating techniques and data from physics, mathematics, and geology, seismology has produced a remarkably sharp picture of the earth's interior that is a primary datum for studying the formation and evolution of terrestrial planets. Seismologists have also learned much about the nature of earthquakes and the tectonic processes responsible for them. These studies are not of purely academic interest; seismology is the major tool for earthquake hazard assessment, hydrocarbon exploration, and the peacekeeping role of nuclear test monitoring.

We thus believe that seismology should be part of the education of every solid earth scientist, rather than a specialized course for those whose primary interest is seismology or other branches of geophysics. The subject has much to offer mineralogists or petrologists studying the composition of the earth's interior, students of tectonics interested in processes of the lithosphere, geologists interested in the nature and evolution of the crust, engineers concerned with seismic hazards, and planetologists interested in the evolution of the terrestrial planets. As the earth sciences become increasingly more integrated and interdisciplinary, the advantages of understanding seismology will continue to grow.

Many students have been deterred from the subject because it requires confronting, often for the first time, both the physics of a continuous medium and wave propagation. We view these concerns as manageable. In fact, we believe that seismology is a good way to introduce these topics, because it applies what might otherwise seem abstract ideas. Seismic waves illustrate effects like reflection, refraction, diffraction, and dispersion by using them to study the earth. Earthquakes demonstrate concepts like rigid tectonic plates, stress and strain, and viscous mantle flow. Thus seismology is a natural way to discuss fundamental processes.

Our goal is to introduce key concepts and their application in present research. This twofold goal places several limitations on the text. First, time and space restrictions require a trade-off between the range of topics and the level of presentation. The resulting choices are, of necessity, subjective. Second, we end discussions when material, however fascinating, seems more appropriate for advanced classes or courses in a related field.[1] Third, these limitations preclude an account of the historical development of the subject, or a systematic assignment of credit for ideas and results. Fourth, in introducing topics of current research, we try to give our sense of issues while recognizing that others' views may differ. The danger in presenting the "current state of knowledge" in a text is that the field changes so rapidly that accounts can soon be out of date. We thus try to focus not on "what we know," but on "how we seek to find out," and highlight current findings in the context of studying interesting questions.

Given these limitations, suggestions for further reading are provided. When possible, the readings are texts or reviews rather than specialized research papers. In many cases, the sources of the figures used to illustrate a concept provide additional information. We also give some references to sites on the World Wide Web, recognizing the trade-off between the wealth of information there and the fact that the Web is volatile and sites can change locations or vanish.

The material is designed for advanced undergraduates and first-year graduate students. Readers are assumed to be familiar with ordinary differential equations and introductory physics. Further background, including basic earth science courses, is helpful but not essential. Material beyond this level is derived as needed. Thus, we seek a balance between presenting the mathematics like magic pulled from a hat and deriving so much so that the thematic flow is disrupted. Hence we

---

[1] Because subfields in the earth sciences overlap, the divisions between them are not sharp, and a given topic draws on several. As John Muir, an early member of the Seismological Society of America better known for founding the Sierra Club, pointed out, "when we look at anything in isolation we realize it is hitched to the rest of the universe."

review some useful mathematics in an Appendix, to which we refer. Other mathematical concepts, notably topics in Fourier analysis, are used as needed and then presented in more depth when appropriate.

Our goal is to introduce some concepts about seismology and its application to such studies of earth structure and earthquakes. Doing this requires developing basic ideas about wave propagation in a continuous solid medium, so the material of greatest interest to geologically oriented readers is somewhat postponed. Readers are urged to enjoy rather than endure the introductory material on elasticity and wave propagation. They risk only discovering the appeal of these topics and finding themselves taking subsequent advanced courses.

Part of the delights of the earth sciences is that they are less structured than some other sciences. There is no single set of topics covered in specific courses, which instead reflect the instructor's and students' interests. Certainly this is the case here. The topics we have chosen contain about a year's worth of class material, which we ourselves divide into several courses. Many students, of course, take only one. We have experimented with different groupings, all of which seemed to work well. We usually do not cover the Appendix in lectures, but assign its problems to identify areas for study or review.

We have found that the homework problems are helpful for understanding the topics. Given the nature of the modern earth sciences, many problems are designed to be done on computers. In our teaching, we expect that most will be done by writing programs, and hence require programming, beginning with simple problems in the Appendix and building to more complex ones in the chapters. A secondary motive is to ensure that students learn the skills of scientific programming, which are often not stressed in computer classes. Some of the problems can be done using spreadsheets, and most can be done with specialized mathematical software.

Some matters of style are worth mentioning. We illustrate interconnections between topics by referring both forward and backward to other sections. Figures are labeled with hyphens (e.g. 5.6-2), and equations with periods (e.g. 5.3.2). Footnotes generally cover side observations which we note in class but are not essential. We use both SI units (those based on the meter, kilogram, and second) and cgs units (those based on the centimeter, gram, and second) because both are common in the literature, although SI units are slowly superseding cgs. We also use other units when customary: seismic velocities are given in km/s and plate motions are given in the more intuitive mm/yr (e.g., 48 mm/yr rather than 1.5×10^{-9} m/s), following Emerson's dictum that "a foolish consistency is the hobgoblin of little minds."

We have enjoyed writing this book. It is a pleasure to try to summarize this diverse and fascinating discipline. We hope readers have as much fun as we did, and that our discussions prompt them to raise interesting and provocative questions as well as learn the material. We also hope that some readers are motivated to continue study of and research on these topics. Much remains to be learned about the earth and earthquake processes, and the opportunities for contributions are great for those with the energy and imagination to go beyond our current knowledge and ideas. Three hundred years after Isaac Newton's work in mechanics and optics laid what would become seismology's foundations, it is worth recalling his words: "I seem to have been only like a boy playing on the seashore, and diverting myself in now and then finding a smoother pebble or a prettier shell than ordinary, whilst the great ocean of truth lay all still undiscovered before me."

Acknowledgments

This book has evolved over many years, with much assistance from students, colleagues, and many who started off as one and became the other. Any effort to thank them all will omit some who should be credited, but is better than none. Similarly, we received much good advice, although not all of it could be accommodated, given the practical issues of manuscript length and level of material.

The text is better for thoughtful questions and assistance from students whose courses used its preliminary forms. In particular, Gary Acton, Don Argus, Craig Bina, John Brodholt, Po-Fei Chen, John DeLaughter, Charles DeMets, George Helffrich, Eryn Klosko, Lisa Leffler, Paul Lundgren, Frederick Marton, Andrew Michael, Andrew Newman, Phillip Richardson, Thomas Shoberg, Paul Stoddard, John Werner, Dale Woods, and Mark Woods helped in developing this material. Many of the figures result from their assistance, as well as the artistic talents of Ranjini Mahinda and Megan Murphy. Cheril Cheverton and Will Kazmeier provided valuable help in manuscript preparation.

We also benefited from suggestions and assistance by colleagues, especially Craig Bina, Raymon Brown, Wang-Ping Chen, Ken Creager, Robert Crosson, Joseph Engeln, Edward Flinn, Yoshio Fukao, Robert Geller, William Holt, Stephen Kirby, Simon McClusky, Emile Okal, Gary Pavlis, Aristeo Pelayo, Steve Roecker, Giovanni Sella, Tetsuzo Seno, Anne Sheehan, Zhang-Kang Shen, Robert Smalley, Robert Smith, Carol Stein, John Vidale, and Douglas Wiens. Jean van Altena, Cameron Laux, John Staples, and Nancy Duffy of Blackwell provided invaluable help.

Indirect support was provided by the National Science Foundation through PECASE award #NSF–EAR–9629018.

Most crucially, Carol Stein and Joan Wysession encouraged the project and allowed us the time required.

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.

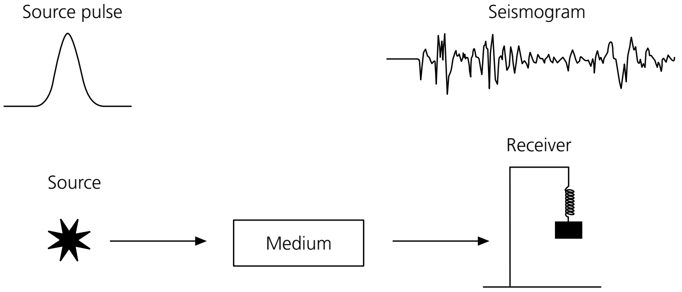

Fig. 1.1-1 Schematic geometry of a seismic experiment.

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.

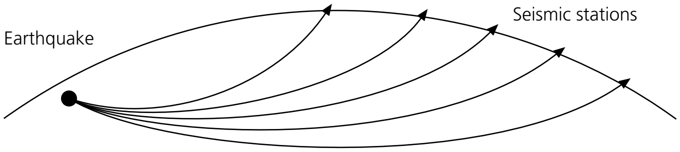

Fig. 1.1-2 Seismic ray paths in the earth, showing the effect of an increase in seismic velocity with increasing depth. The waves travel in curved paths between the earthquake and seismic stations.

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.

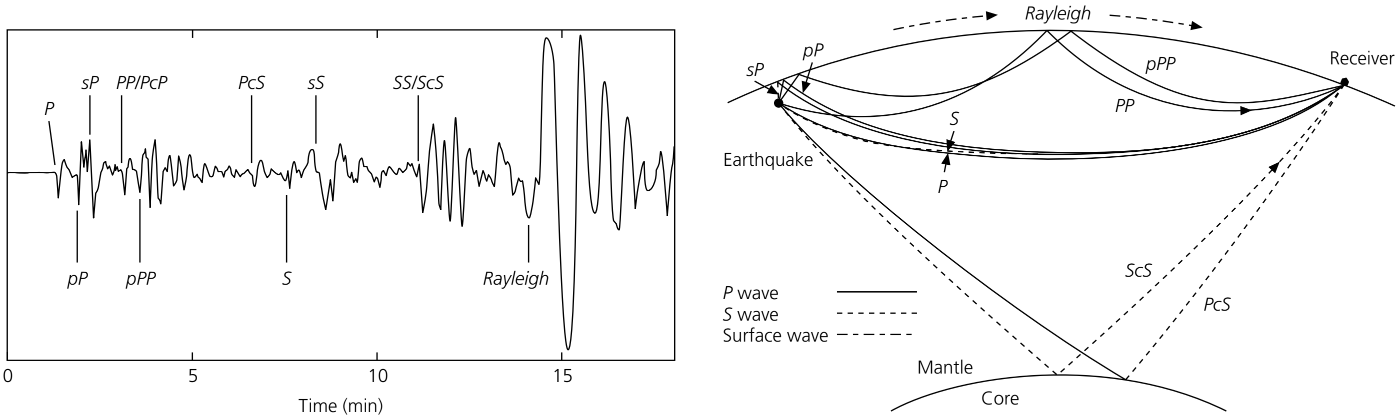

Fig. 1.1-3 *Left*: Long-period vertical component seismogram at Golden, Colorado, from an earthquake in Colombia (July 29, 1967), showing various seismic phases. The distance from earthquake to station is 44°. *Right*: Ray paths for the seismic phases labeled on the seismogram.

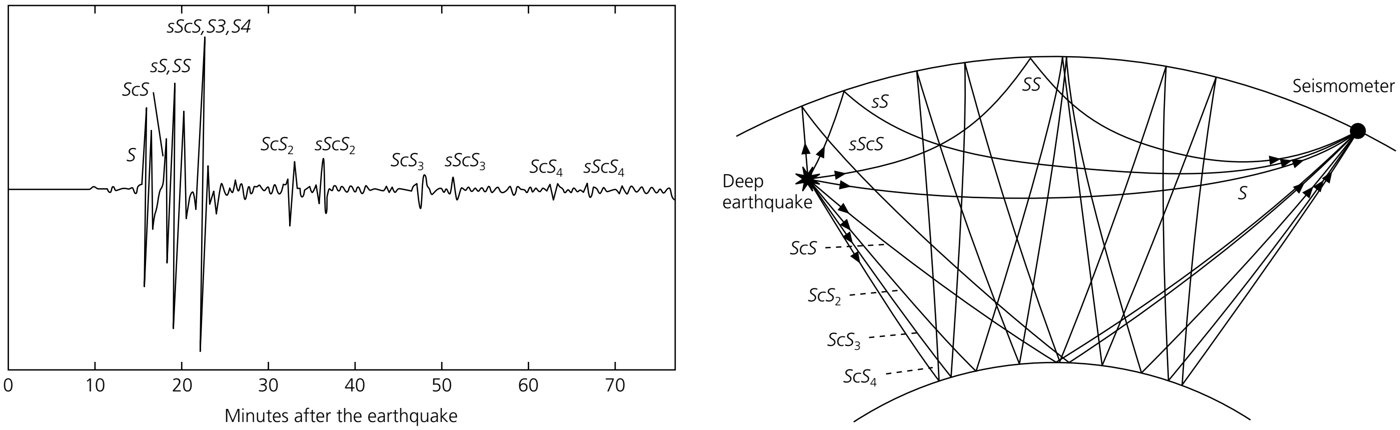

Fig. 1.1-4 Seismogram (*left*) and ray paths (*right*) for a deep focus earthquake in Tonga, recorded at Oahu (Hawaii), showing multiple core reflections.

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

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

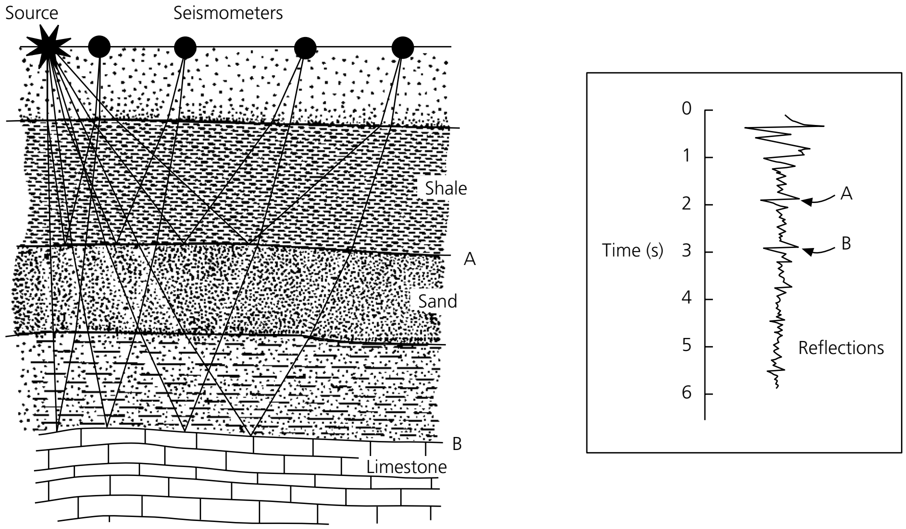

Fig. 1.1-5 Schematic example of the seismic reflection method, the basic tool of hydrocarbon exploration.

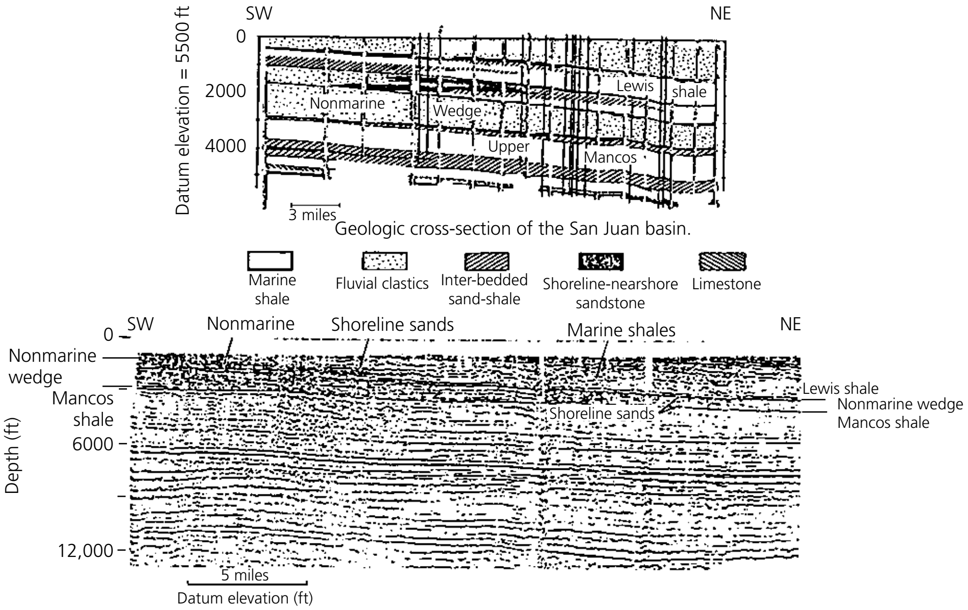

Fig. 1.1-6 Data from a reflection seismic survey across the San Juan Basin, New Mexico (*bottom*) and the resulting geological interpretation (*top*). [Sangree and Widmier, 1979. Reprinted by permission of the Society of Exploration Geophysicists.)

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

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[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.

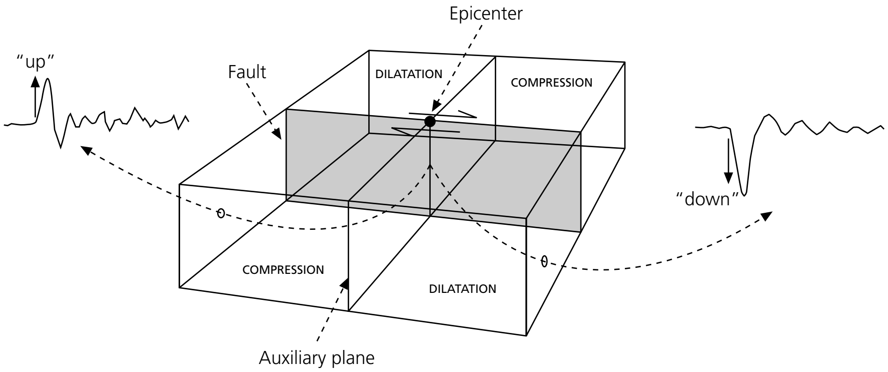

Fig. 1.1-7 First motions of seismic *P* waves observed at seismometers located in various directions about the earthquake allow the fault orientation to be determined.

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

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[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

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[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.

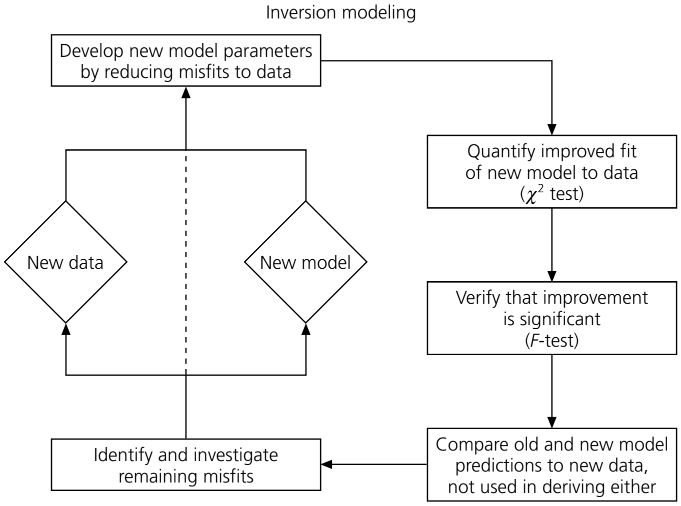

Fig. 1.1-8 Schematic illustration of how models of earth processes advance with time due to additional data and improved model parameterizations.

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

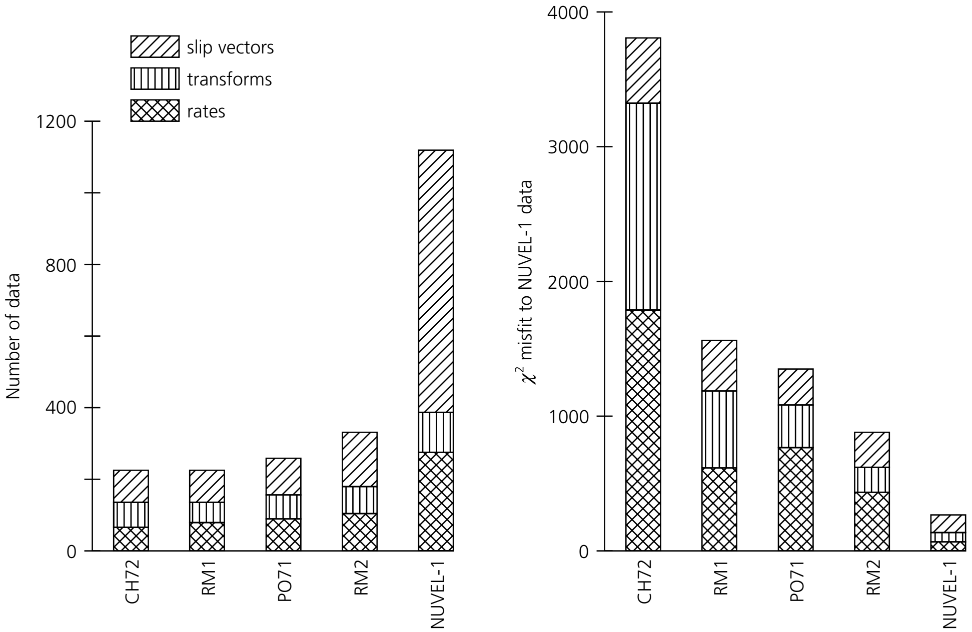

Fig. 1.1-9 Evolution of successive global plate motion models, as the amount of data increases and the misfit is reduced. *Left*: Number of data used to derive the models. Three types of data are inverted: earthquake slip vector azimuths, transform fault azimuths, and spreading rates. *Right*: The misfit 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.)

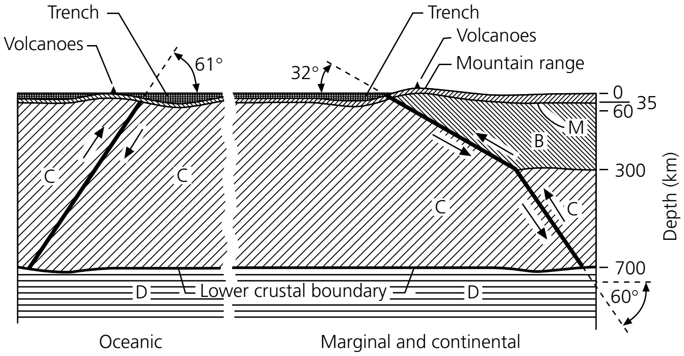

Fig. 1.1-10 Tectonic cartoon for oceanic and continental margin trenches, prior to the acceptance of plate tectonics. The association of dip-slip earthquakes with trenches, volcanism, and mountain ranges was recognized. Note the exaggeration of surface relief. (Benioff, 1955. From *Crust of the Earth*, 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.

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]

[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] 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."

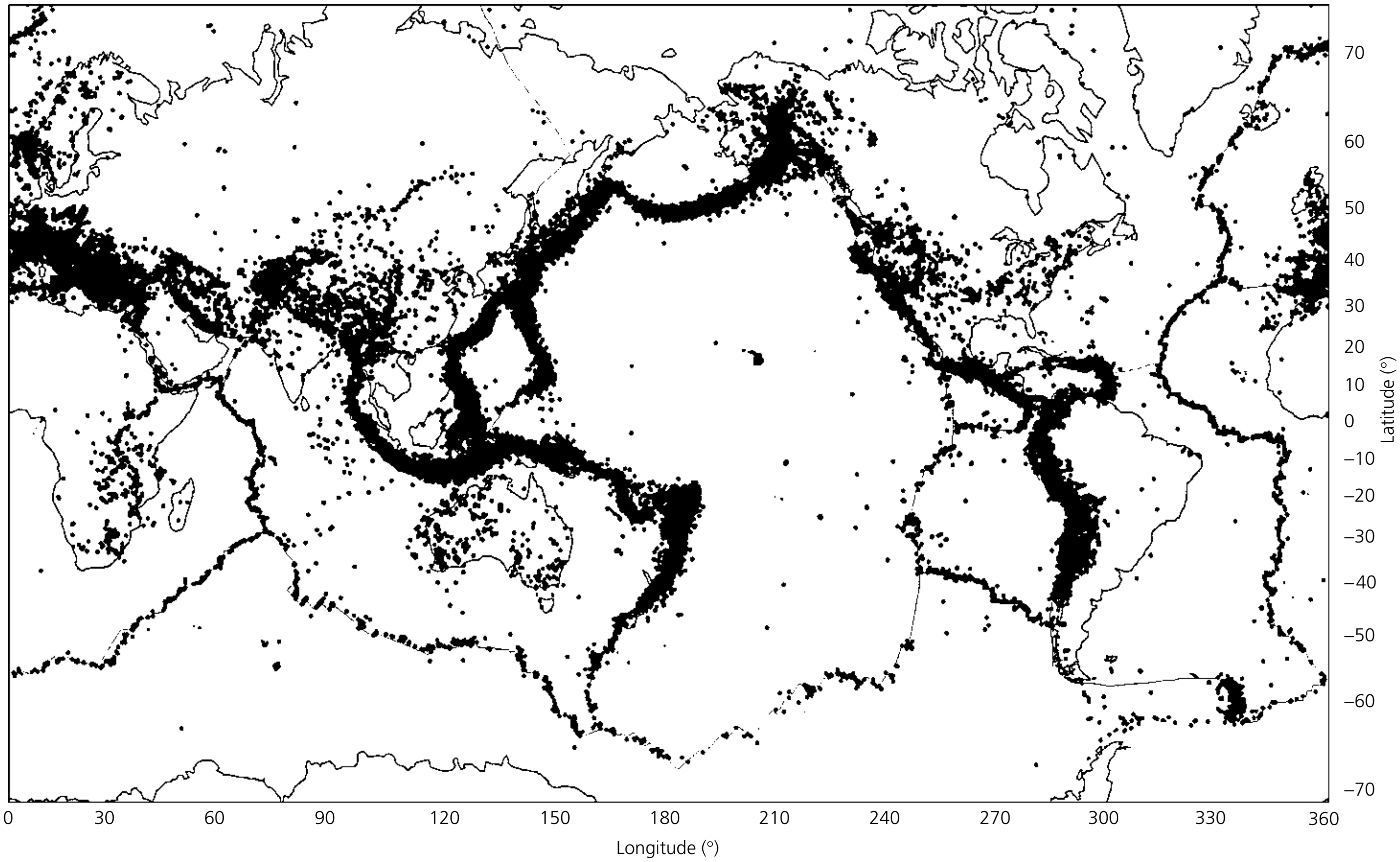

Fig. 1.2-1 Map showing epicenters of all earthquakes during 1963–95 with magnitudes of $m_b \geq 4$. Most earthquakes occur along the boundaries between tectonic plates. Where these boundaries are distinct, the earthquakes occur within narrow bounds. More diffuse plate boundaries, like the Himalayan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

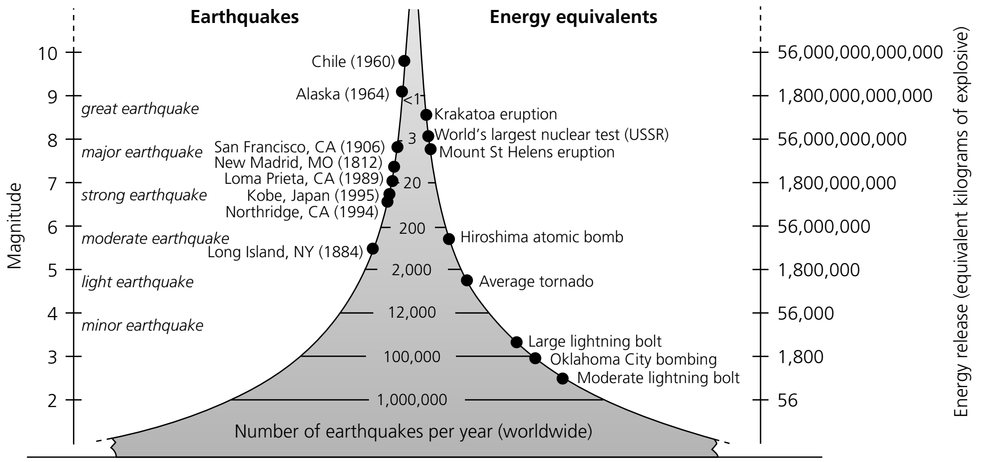

Fig. 1.2-2 Comparison of frequency, magnitude, and energy release of earthquakes and other phenomena. The magnitude used is moment magnitude, M_w. (After Incorporated Research Institutions for Seismology.)

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

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[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

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.

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[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.

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 ±

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[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.

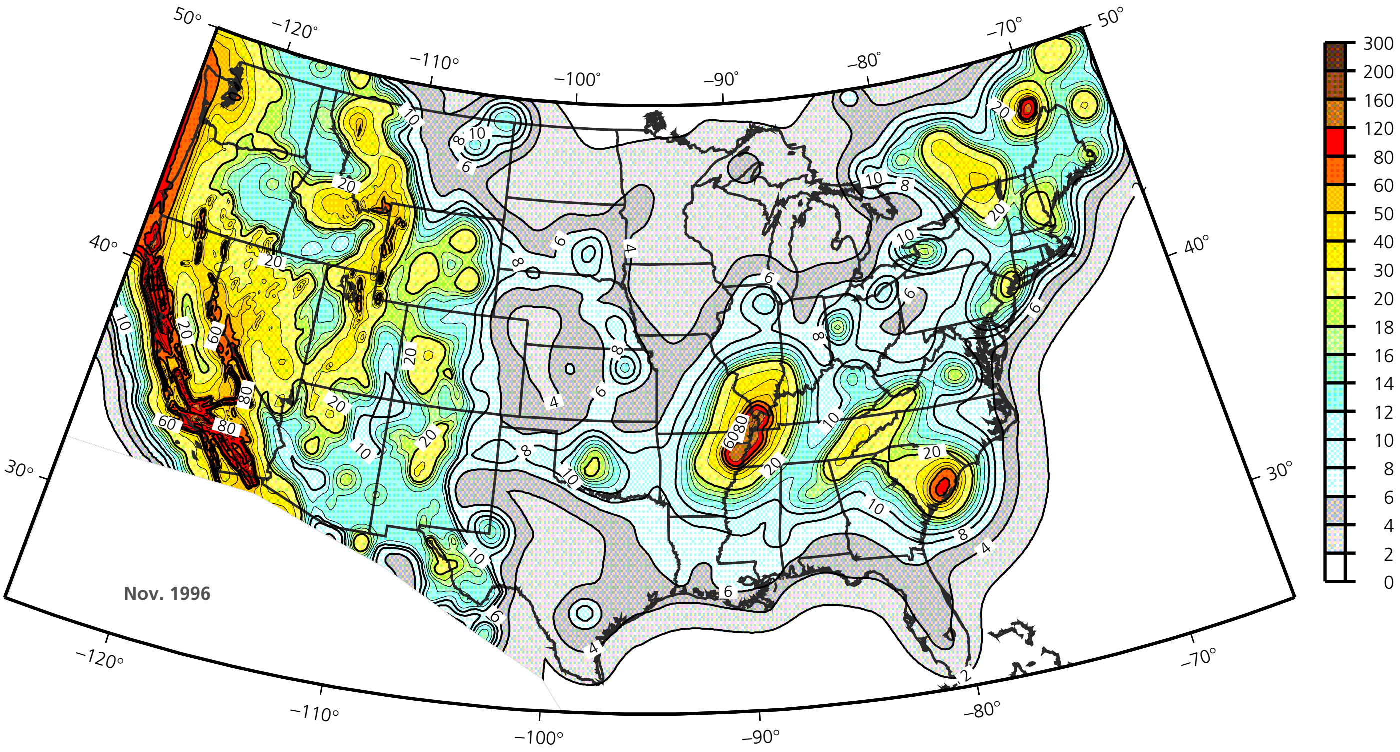

Fig. 1.2-3 A map of estimated earthquake hazards in the United States. The predicted hazards are plotted as the maximum acceleration of ground shaking expected at a 2% probability over a 50-year period. Although the only active plate boundaries are in the western USA, other areas are also shown as having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

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[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).

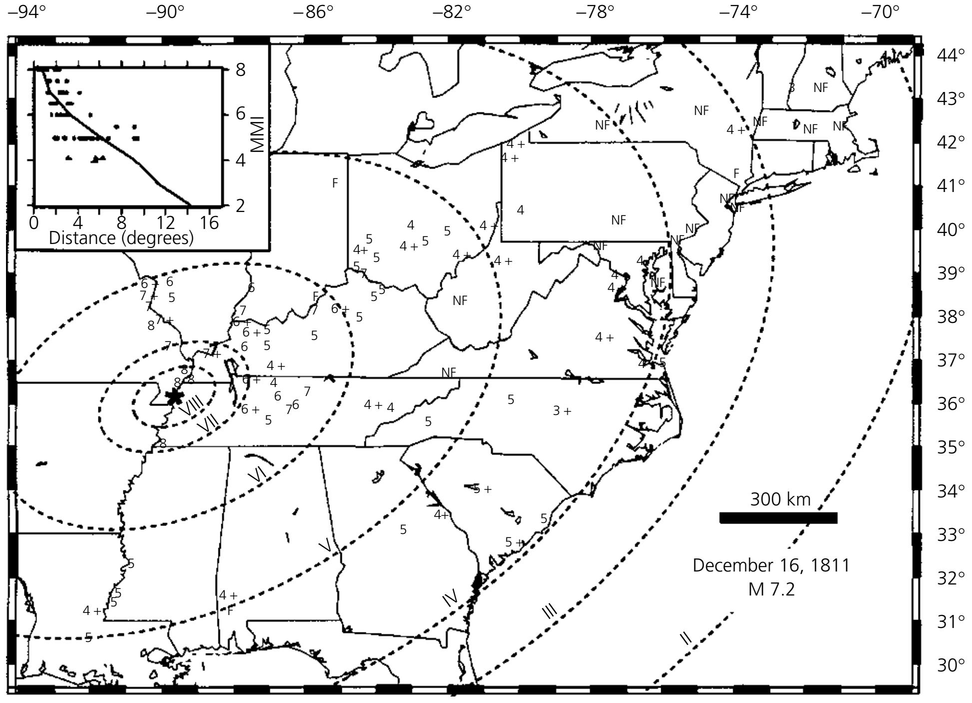

Fig. 1.2-4 Isoseismals for the first of the three largest earthquakes of the 1811–12 New Madrid earthquake sequence. Such plots, though based on sparse data, often provide the best assessment of historical earthquakes and of the effects of future ones. (After Hough *et al.*, 2000. *J. Geophys. Res.*,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