Earthquake Monitoring and Early Warning Systems

WILLIAM H. K. LEE[1], YIH-MIN WU[2]
[1] US Geological Survey, Menlo Park, USA
[2] Department of Geosciences, National Taiwan University, Taipei, Taiwan

Article Outline

Glossary

Active fault A *fault* (q.v.) that has moved in historic (e. g., past 10,000 years) or recent geological time (e. g., past 500,000 years).

Body waves Waves which propagate through the interior of a body. For the Earth, there are two types of seismic body waves: (1) compressional or longitudinal (*P* wave), and (2) shear or transverse (*S* wave).

Coda waves Waves which are recorded on a *seismogram* (q.v.) after the passage of *body waves* (q.v.) and *surface waves* (q.v.). They are thought to be back-scattered waves due to the Earth's inhomogeneities.

Earthquake early warning system (EEWS) An earthquake monitoring system that is capable of issuing warning message after an earthquake occurred and before strong ground shaking begins.

Earthquake precursor Anomalous phenomenon preceding an earthquake.

Earthquake prediction A statement, in advance of the event, of the time, location, and *magnitude* (q.v.) of a future earthquake.

Epicenter The point on the Earth's surface vertically above the *hypocenter* (q.v.).

Far-field Observations made at large distances from the *hypocenter* (q.v.), compared to the wave-length and/or the source dimension.

Fault A fracture or fracture zone in the Earth along which the two sides have been displaced relative to one another parallel to the fracture.

Fault slip The relative displacement of points on opposite sides of a *fault* (q.v.), measured on the fault surface.

Focal mechanism A description of the orientation and sense of slip on the causative fault plane derived from analysis of *seismic waves* (q.v.).

Hypocenter Point in the Earth where the rupture of the rocks originates during an earthquake and *seismic waves* (q.v.) begin to radiate. Its position is usually determined from arrival times of *seismic waves* (q.v.) recorded by *seismographs* (q.v.).

Intensity, earthquake Rating of the effects of earthquake vibrations at a specific place. Intensity can be estimated from instrumental measurements, however, it is formally a rating assigned by an observer of these effects using a descriptive scale. Intensity grades are commonly given in Roman numerals (in the case of the Modified Mercalli Intensity Scale, from I for "not perceptible" to XII for "total destruction").

Magnitude, earthquake Quantity intended to measure the size of earthquake at its source, independent of the place of observation. *Richter magnitude* (M_L) was originally defined in 1935 as the logarithm of the maximum amplitude of seismic waves in a seismogram written by a Wood–Anderson seismograph (corrected to) a distance of 100 km from the epicenter. Many types of magnitudes exist, such as *body-wave magnitude* (m_b), *surface-wave magnitude* (M_S), and *moment magnitude* (M_W).

Moment tensor A symmetric second-order tensor that characterizes an internal seismic point source completely. For a finite source, it represents a point source approximation and can be determined from the analysis of *seismic waves* (q.v.) whose wavelengths are much greater than the source dimensions.

Near-field A term for the area near the causative rupture of an earthquake, often taken as extending a distance from the rupture equal to its length. It is also used to specify a distance to a seismic source comparable or shorter than the wavelength concerned. In engineering applications, near-field is often defined as the area within 25 km of the fault rupture.

Plate tectonics A theory of global *tectonics* (q.v.) in which the Earth's lithosphere is divided into a number of essentially rigid plates. These plates are in relative motion, causing earthquakes and deformation along the plate boundaries and adjacent regions.

Probabilistic seismic hazard analysis Available information on earthquake sources in a given region is combined with theoretical and empirical relations among earthquake *magnitude* (q.v.), distance from the source, and local site conditions to evaluate the exceedance probability of a certain ground motion parameter, such as the peak ground acceleration, at a given site during a prescribed time period.

Seismic hazard Any physical phenomena associated with an earthquake (e. g., ground motion, ground failure, liquefaction, and tsunami) and their effects on land use, man-made structure, and socio-economic systems that have the potential to produce a loss.

Seismic hazard analysis The calculation of the *seismic hazard* (q.v.), expressed in probabilistic terms (See *probabilistic seismic hazard analysis*, q.v.). The result is usually displayed in a *seismic hazard map* (q.v.).

Seismic hazard map A map showing contours of a specified ground-motion parameter or response spectrum ordinate for a given *probabilistic seismic hazard* analysis (q.v.) or return period.

Seismic moment The magnitude of the component couple of the double couple that is the point force system equivalent to a *fault slip* (q.v.) in an isotropic elastic body. It is equal to rigidity times the fault slip integrated over the fault plane. It can be estimated from the far-field seismic spectrum at wave lengths much longer than the source size. It can also be estimated from the near-field seismic, geologic and geodetic data. Also called "scalar seismic moment" to distinguish it from *moment tensor* (q.v.).

Seismic risk The risk to life and property from earthquakes.

Seismic wave A general term for waves generated by earthquakes or explosions. There are many types of seismic waves. The principle ones are *body waves* (q.v.), *surface waves* (q.v.), and *coda waves* (q.v.).

Seismograph Instrument which detects and records ground motion (and especially vibrations due to earthquakes) along with timing information. It consists of a *seismometer* (q.v.) a precise timing device, and a recording unit (often including telemetry).

Seismogram Record of ground motions made by a *seismograph* (q.v.).

Seismometer Inertial sensor which responds to ground motions and produces a signal that can be recorded.

Source parameters of an earthquake The parameters specified for an earthquake source depends on the assumed earthquake model. They are origin time, *hypocenter* (q.v.), *magnitude* (q.v.), *focal mechanism* (q.v.), and *moment tensor* (q.v.) for a point source model. They include fault geometry, rupture velocity, stress drop, slip distribution, etc. for a finite fault model.

Surface waves Waves which propagate along the surface of a body or along a subsurface interface. For the Earth, there are two common types of seismic surface waves: Rayleigh waves and Love waves (both named after their discoverers).

Tectonics Branch of Earth science which deals with the structure, evolution, and relative motion of the outer part of the Earth, the lithosphere. The lithosphere includes the Earth's crust and part of the Earth's upper mantle and averages about 100 km thick. See *plate tectonics* (q.v.).

Teleseism An earthquake at an epicentral distance greater than about 20° or 2000 km from the place of observation.

Definition of the Subject

When a sudden rupture occurs in the Earth, elastic (seismic) waves are generated. When these waves reach the Earth's surface, we may feel them as a series of vibrations, which we call an earthquake. *Seismology* is derived from the Greek word $\sigma\epsilon\iota\sigma\mu\acute{o}\varsigma$ (seismos or earthquake) and $\lambda\acute{o}\gamma o\varsigma$ (logos or discourse); thus, it is the science of earthquakes and related phenomena. Seismic waves can be generated naturally by earthquakes or artificially by explosions or other means. We define earthquake monitoring as a branch of seismology, which systematically observes earthquakes with instruments over a long period of time.

Instrumental recordings of earthquakes have been made since the later part of the 19th century by seismographic stations and networks of various sizes from local to global scales. The observed data have been used, for example, (1) to compute the source parameters of earthquakes, (2) to determine the physical properties of the Earth's interior, (3) to test the theory of plate tectonics, (4) to map active faults, (5) to infer the nature of damaging ground shaking, and (6) to carry out seismic hazard analyzes. Constructing a satisfactory theory of the complex earthquake process has not yet been achieved within the context of physical laws, e. g., realistic equations for modeling earthquakes do not exist at present. Good progress, however, has been made in building a physical foundation for the earthquake source process [62], partly as a result of research directed toward earthquake prediction.

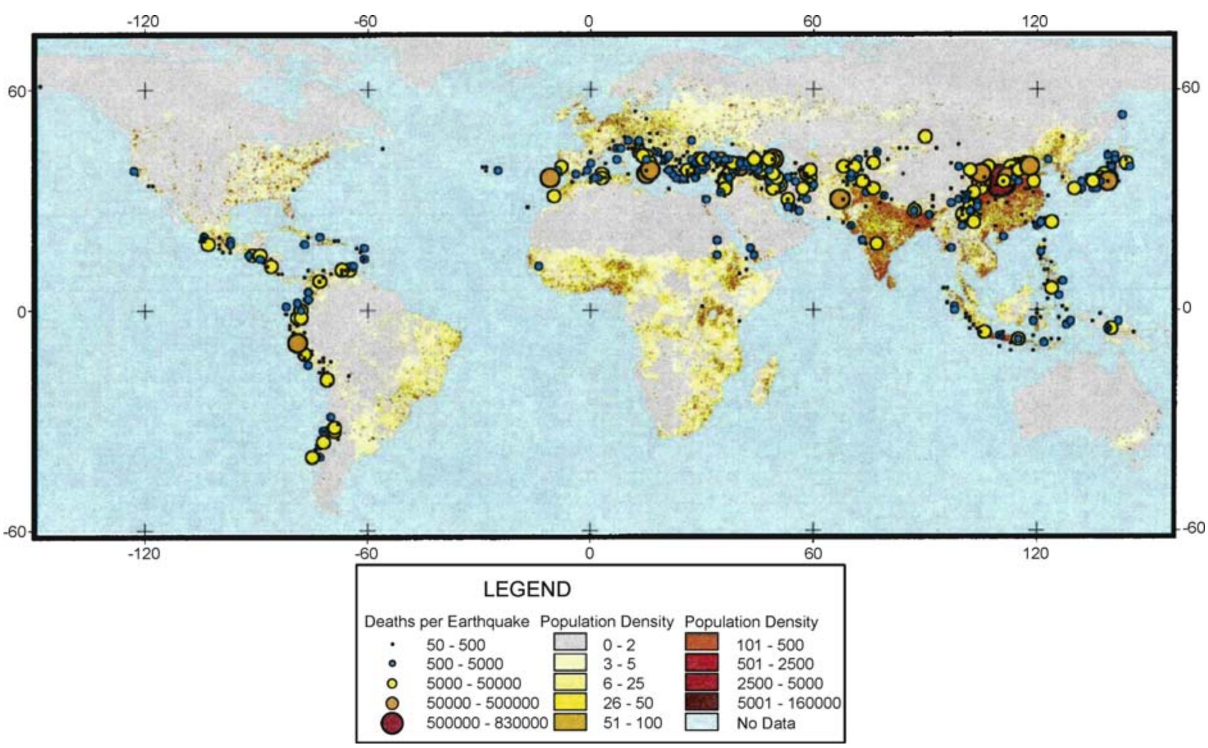
Earthquakes release large amounts of energy that potentially can cause significant damage and human deaths. During an earthquake, potential energy (mainly elastic strain energy and some gravitational energy) that has accumulated in the hypocentral region over decades to centuries or longer is released suddenly [63]. This energy is partitioned into (1) radiated energy in the form of propagating seismic waves, (2) energy consumed in overcoming fault friction, (3) the energy which expands the rupture surface area or changes its properties (e. g., by pulverizing rock), and (4) heat. The radiated seismic energy is a small fraction (about 7%) of the total energy budget, and it can be estimated using the recorded seismograms. Take, for example, the 1971 San Fernando earthquake ($M_W = 6.6$) in southern California. Its radiated energy was about 5×10^{21} ergs, or about 120 kilotons of TNT explosives, or the energy released by six atomic bombs of the size used in World War II. The largest earthquake recorded instrumentally (so far) is the 1960 Chilean earthquake ($M_W = 9.5$). Its radiated energy was about 1.1×10^{26} ergs, an equivalent of about 2,600 megatons of TNT explosives, the energy released by about 130,000 atomic bombs. It is, therefore, no surprise that an earthquake can cause up to hundreds of thousands of human deaths, and produce economic losses of up to hundreds of billions of dollars.

Monitoring earthquakes is essential for providing scientific data to investigate complex earthquake phenomena, and to mitigate seismic hazards. The present article is a brief overview of earthquake monitoring and early warning systems; it is intended for a general scientific audience, and technical details can be found in the cited references. Earthquakes are complex natural phenomena and their monitoring requires an interdisciplinary approach, including using tools from computer science, electrical and electronic engineering, mathematics, physics, and others. Earthquake early warning systems (which are based on earthquake monitoring) offer practical information for reducing seismic hazards in earthquake-prone regions.

After the "Introduction", we will present a summary of earthquake monitoring, a description of the products derived from the analysis of seismograms, and a discussion of the limitations of these products. Earthquake early warning systems are then presented briefly, and we conclude with a section on future directions, including a progress report on rotational seismology (Appendix). We present overviews of most topics in earthquake monitoring, and an extensive bibliography is provided for additional reading and technical details.

Introduction

Earthquakes, both directly and indirectly, have caused much suffering to mankind. During the 20th century alone about two million people were killed as a result of earthquakes. A list of deadly earthquakes (death tolls ≥ 25) of the world during the past five centuries was compiled by Utsu [115]. It shows that earthquakes of magnitude ≥ 6 (~ 150 per year worldwide) can be damaging and deadly if they occur in populated areas, and if their focal depths are shallow (< 50 km). Seismic risk can be illustrated by plotting the most deadly earthquakes of the past five centuries (1500–2000) over a map of current population density. This approach was used by Utsu [115], and his result is shown in Fig. 1. Most of these deadly earthquakes are concentrated (1) along the coasts of Central America, the Caribbean, western South America, and Indonesia, and (2) along a belt that extends from southern Europe,



Earthquake Monitoring and Early Warning Systems, Figure 1
Location of deadly earthquakes around the world, 1500–2000. Population density is shown by the *background colors*. See [115] for details

Earthquake Monitoring and Early Warning Systems, Table 1
Deadly Earthquakes/Tsunamis from 1896–2005 ([115] and recent sources)

Origin Time Year MM/DD Hr:Min (UTC, except L=local)	Hypocenter Lat. (deg)	Lon. (deg)	Depth (km)	Magnitude	Location	Deaths (Approximate)
2005 10/08 3:50	34.432	73.573	10	7.6	Pakistan, Kashmir	80,361+
2004 12/26 0:58	3.298	95.778	7	9.2	Indonesia, Sumatra	283,106+
2003 12/26 1:56	29.004	58.337	15	6.6	Iran, Bam	26,000
2001 01/26 3:16	23.420	70.230	16	7.7	India, Gujarat, Bhuj	20,000+
1990 06/20 21:00	37.008	49.213	18	7.4	Iran, western	∼40,000
1988 12/07 7:41	40.919	44.119	7	6.8	Armenia, Spitak	∼40,000
1976 07/27 19:42	39.605	117.889	17	7.6	China, Tangshan	∼242,000
1976 02/04 9:01	15.298	−89.145	13	7.5	Guatemala	23,000
1970 05/31 20:23	−9.248	−78.842	73	7.5	Peru	67,000
1948 10/05 20:12	37.500	58.000	0	7.2	USSR, Ashgabat	∼65,000
1939 12/26 23:57	39.770	39.533	35	7.7	Turkey, Erzincan	33,000
1939 01/25 3:32	−36.200	−72.200	0	7.7	Chile, Chillian	28,000
1935 05/30 21:32	28.894	66.176	35	8.1	Pakistan, Quetta	60,000
1932 12/25 2:04	39.771	96.690	25	7.6	China, Gansu	∼70,000
1927 05/22 22:32	37.386	102.311	25	7.7	China, Tsinghai	∼100,000
1923 09/01 2:58	35.405	139.084	35	7.9	Japan, Kanto	143,000
1920 12/16 12:05	36.601	105.317	25	8.6	China, Gansu	∼240,000
1915 01/13 6:52	42.000	13.500	0	6.9	Italy, Avezzano	33,000
1908 12/28 4:20	38.000	15.500	0	7.0	Italy, Messina	∼ 82,000
1906 08/17 0:40	−33.000	−72.000	0	8.2	Chile, Valparaiso	20,000
1905 04/04 0:50	33.000	76.000	0	8.1	India, Kangra	20,000
1896 06/15 19:32L	39.500	144.000	0	8.2	Japan, Sanriku-oki	22,000

"∼" denotes large uncertainties because a range of deaths had been reported.
"+" denotes a minimum value.

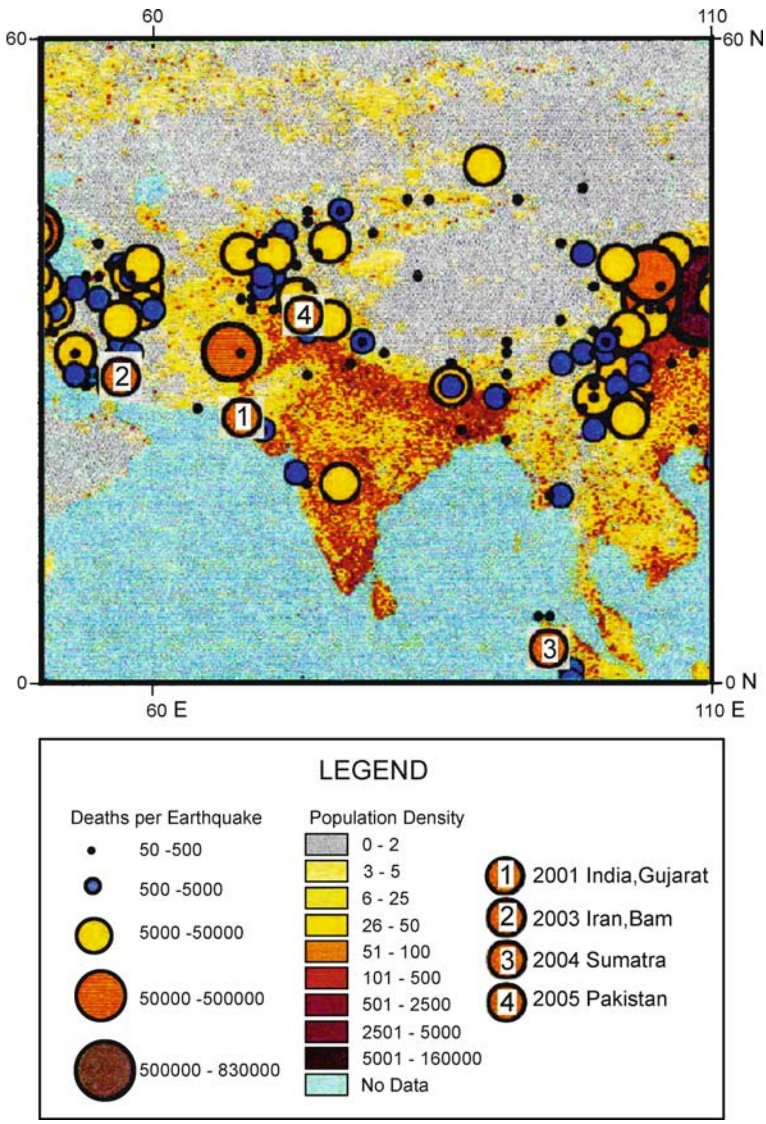
the Middle East, Iran, Pakistan and India, to China and Japan.

Table 1 lists the most deadly earthquakes (death toll $> 20,000$) of the past 110 years based on official estimates (often under-estimated for political reasons, or lack of accurate census data in many areas of the world). In the first 5 years of the 21st century, four disastrous earthquakes occurred in India, Indonesia, Iran, and Pakistan. In the 20th century, the average death toll caused by earthquakes (and tsunamis they triggered) was about 16,000 per year. For the past seven years the yearly death toll was about 60,000 – four times higher than the average in the previous century. In Fig. 2 we extracted a portion of Fig. 1 to illustrate the relationship between past earthquakes and population in India, Pakistan, northern Indonesia, and adjoining regions. We numbered the four most recent disastrous earthquakes in Fig. 2. It is obvious that the large populations in India, Indonesia, Iran, Pakistan, and their adjoining regions (over 1.5 billion people) has been and will continue to be adversely affected by earthquakes. Fatalities depend largely on resistance of building construction to shaking, in addition to population density and earthquake occurrence.

In recent decades, population increases, accelerated urbanization, and population concentration along coastal areas prone to earthquakes suggest that many more earthquake-related fatalities will occur unless effective steps are taken to minimize earthquake and tsunami hazards.

Earthquake Monitoring: Instrumentation

Besides geodetic data [28], the primary instrumental data for the quantitative study of earthquakes are *seismograms*, records of ground motion caused by the passage of seismic waves. Seismograms are written by *seismographs*, instruments which detect and record ground motion along with timing information. A seismograph consists of three basic components: (1) a seismometer, which responds to ground motion and produces a signal proportional to acceleration, velocity, or displacement over a range of amplitudes and frequencies; (2) a timing device; (3) either a local recording unit which writes seismograms on paper, film, or elec-



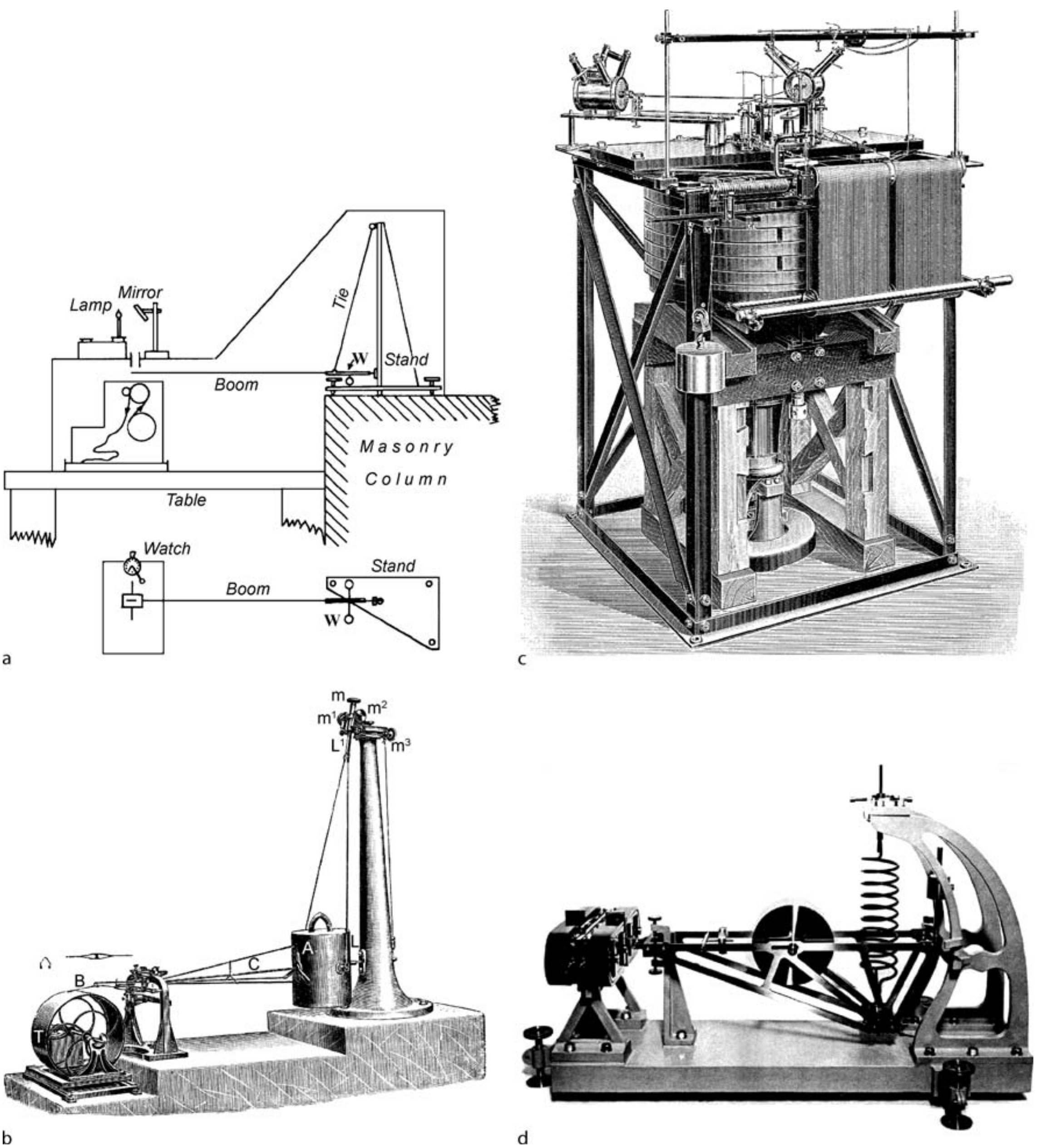
Earthquake Monitoring and Early Warning Systems, Figure 2
Location of the 4 most deadly earthquakes of the 21st century (up to the end of 2007) on a map showing the location of the deadly earthquakes from 16th to 20th centuries (after [115] and Table 1)

tronic storage media, or more recently, a telemetry system for delivering the seismograms to a central laboratory for recording. Technical discussions of seismometry may be found, for example, in Wielandt [122], and of seismic instruments in Havskov and Alguacil [48]. An overview of challenges in observational earthquake seismology is given by Lee [71], and a useful manual of seismological observatory practice is provided by Bormann [12].

An *accelerograph* is a seismograph designed to record, on scale, the acceleration time history of strong ground motions. Measuring acceleration is important for studying response of buildings to strong ground motions close to earthquakes. Many modern sensitive seismographs are *velocigraphs* recording the time history of ground velocity. They are designed to measure seismic waves of small amplitudes (because seismic waves attenuate quickly from their sources) either from small earthquakes nearby, or from large earthquakes that are far away.

A seismic network (or an "array") is a group of seismographs "linked" to a central headquarters. Nowadays the link is by various methods of telemetry, but in early days the links were by mail or telegrams, or simply by manual collection of the records. When we speak of a seismic *station*, we may mean an observatory with multiple instruments in special vaults or a small instrument package at a remote site.

Seismographs were first developed in the late 19th century, and individual seismographic observatories (often a part of astronomical or meteorological observatories) began earthquake monitoring by issuing earthquake information in their station bulletins and other publications. However, in order to accurately locate an earthquake, data from several seismographic stations are necessary. It was then natural for many governments to assume responsibility for monitoring earthquakes within their territories. However, because seismic waves from earthquakes do not



Earthquake Monitoring and Early Warning Systems, Figure 3
Some classical seismographs: a Milne, b Bosch–Omori, c Wiechert, and d Galitzin (after [101])

recognize national boundaries, the need for international cooperation became clear. In the following subsections, we present an overview of the history and results of earthquake monitoring.

Historical Developments

In 1897, John Milne designed the first inexpensive seismograph, which was capable of recording very large earthquakes anywhere in the world. With a small grant from the British Association for the Advancement of Science (BAAS), a few other donations, and his own money, Milne managed to deploy about 30 of his instruments around the world, forming the first worldwide seismographic network. At the same time, seismogram readings were reported voluntarily to Milne's observatory at Shide on the Isle of Wight, England. A global earthquake summary with these seismogram readings was issued by Milne beginning in 1899. These summaries are now known as the "Shide Circulars". Milne also published progress and results in the "Reports of the BAAS Seismological Committee" from 1895 to 1913. A review of Milne's work and a reproduction of his publications as computer readable files were given by Schweitzer and Lee [101] and its attached CD-ROM. After Milne's death in 1913, Herbert H. Turner continued Milne's efforts, and in 1918 established publication of the International Seismological Summary (ISS).

The shortcomings of the Milne seismograph (low magnification, no damping, and poor time resolution) were soon recognized. Several improved seismographs (notably the Omori, Bosch–Omori, Wiechert, Galitzin, and Milne–Shaw) were developed and deployed in the first three decades of the 20th century. Figure 3 shows several of these classical seismographs (see Schweitzer and Lee [101] for further explanation). Although the ISS provided an authoritative compilation arrival-time data of seismic waves and determinations of earthquake hypocenters beginning in 1918, its shortcomings were also evident. These include difficulties in collecting the available arrival-time data around the world (which were submitted on a voluntary basis), and in the processing and analysis of data from many different types of seismographs. Revolutions and wars during the first half of the 20th century frequently disrupted progress, particularly impacting collection and distribution earthquake information.

In the late 1950s, attempts to negotiate a comprehensive nuclear test ban treaty failed, in part because of perceptions that seismic methods were inadequate for monitoring underground nuclear tests [95]. The influential Berkner report of 1959 therefore advocated major support for seismology [66]. As a result, the Worldwide Standardized Seismograph Network (WWSSN) was created in the early 1960s with about 120 continuously recording stations located across much of the world, except China and the USSR [91]. Each WWSSN station was equipped with *identical* sets of short-period and long-period three-component seismographs and accurate chronometers. Figure 4 shows some of the equipment at a WWSSN station, including three-components of long-period seismometers, long-period recording and test instruments, and the time and power console. A similar set of three-component short-period seismometers and recording and test instruments, nearly identical in appearance, was also deployed at each station. Seismograms from the WWSSN were sent to the United States to be photographed on 70 mm film chips for distribution (about US$ 1 per chip as then sold to any interested person).

The WWSSN network is credited with making possible rapid progress in global seismology, and with helping to spark the plate tectonics revolution of the late 1960s [117]. At about the same time, the Unified System of Seismic Observations (ESSN) of the former USSR and its allied countries was established, consisting of almost 100 stations equipped with Kirnos short-period, 1–20 s displacement sensors, and long-period seismographs.

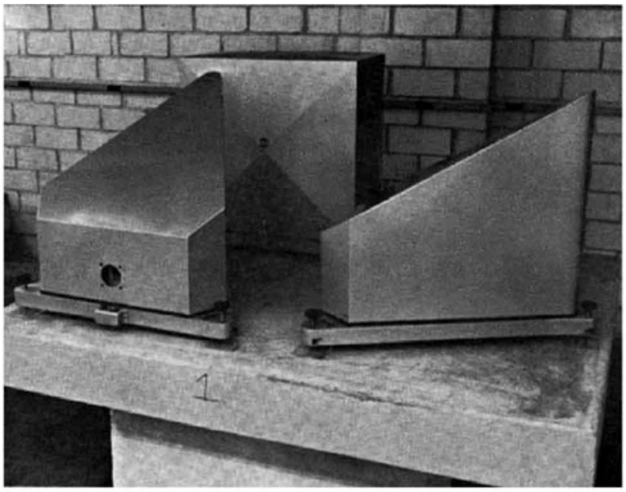
Samples of seismograms recorded on smoked paper and photographic paper or film by analog seismographs are shown in Figs. 5 and 6. Two efforts to preserve and make such records available online are now underway: the *SeismoArchives* (www.iris.edu/seismo/ [72]), and *Sismos* (sismos.rm.ingv.it [82]).

With the establishment of the WWSSN, the United States also assumed the task of monitoring earthquakes on a global scale beginning in the early 1960s. The mission of the US National Earthquake Information Center (NEIC, now part of the US, Geological Survey) is "to determine rapidly the location and size of all destructive earthquakes worldwide and to immediately disseminate this information to concerned national and international agencies, scientists, and the general public" (http://earthquake.usgs.gov/regional/neic/).

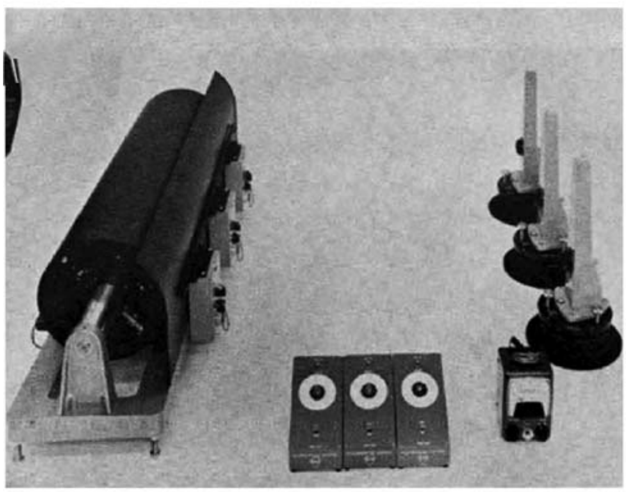
In 1964, the ISS was reorganized as the International Seismological Centre (ISC). Since then, the ISC (http://www.isc.ac.uk/) has issued annual global earthquake catalogs with a time lag of about two years [123].

Technical Considerations

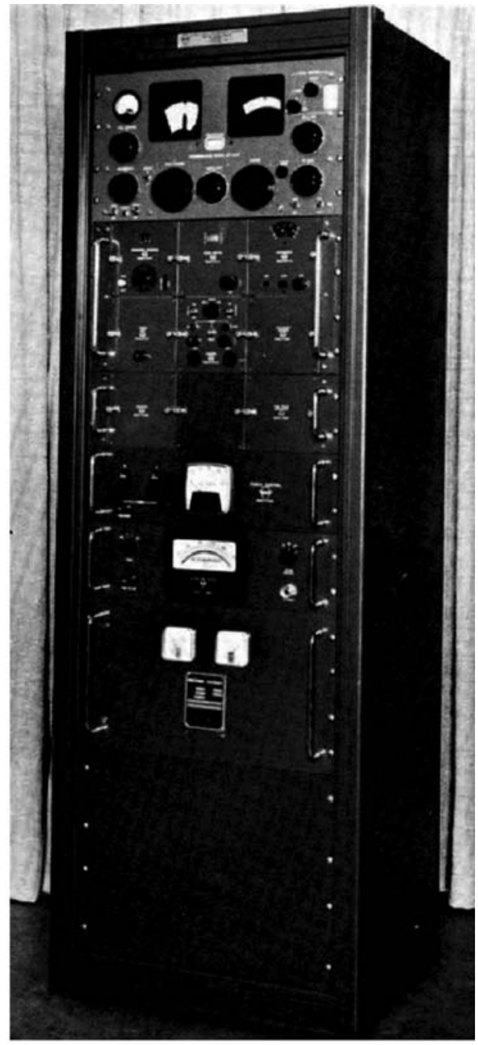
To record seismic waves, we must consider both the available technology for designing seismographs, and the nature of the Earth's background noise [121]. The Earth is constantly in motion. This "background" noise is usu-



a LONG-PERIOD SEISMOMETERS INSTALLED ON A PIER



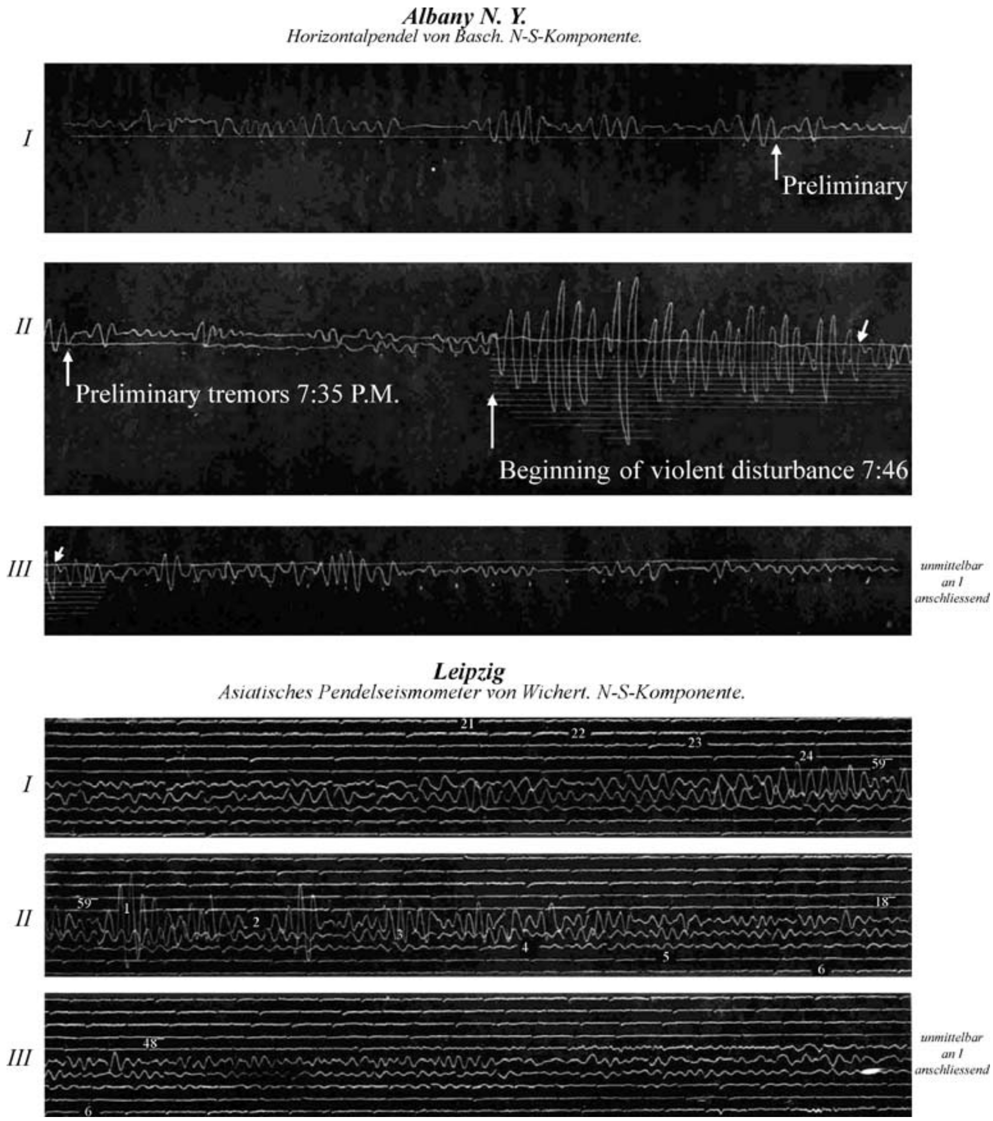
b LONG-PERIOD AND TEST INSTRUMENTS



c TIME AND POWER CONSOLE

Earthquake Monitoring and Early Warning Systems, Figure 4
Some WWSSN station equipment: a Three-component, long-period seismometers installed on a seismic pier, b Long-period recording and test instruments, and c Time and power console. A similar set of three-component, short-period seismometers and recording/test instruments is not shown

ally classified as either (1) *microseisms*, which typically have frequencies below about 1 Hz, are often the largest background signals, and are usually caused by natural disturbances (largely caused by ocean waves near shorelines); or (2) *microtremors*, which have frequencies higher than about 1 Hz, and are due to human activities (such as traffic and machinery) and local natural sources (such as wind and moving vegetation). Ground motions from earthquakes vary more than ten orders of magnitude in amplitude and six orders of magnitude in frequency, depending on the size of the earthquake and the distance at which it is recorded. Figure 7 illustrates the relative dynamic range of some common seismometers for global earthquake monitoring. A "low Earth noise" model [10,92] is the lower limit of Earth's natural noise in its quietest locations – it is desirable to have instruments that are sensitive enough to detect this minimal background Earth signal. In the analog instrument era (i. e., prior to about 1980), short-period and long-period seismometers were designed separately to avoid microseisms, which have predominant periods of about 6 s. Short-period seismometers were designed to detect tiny ground motions from

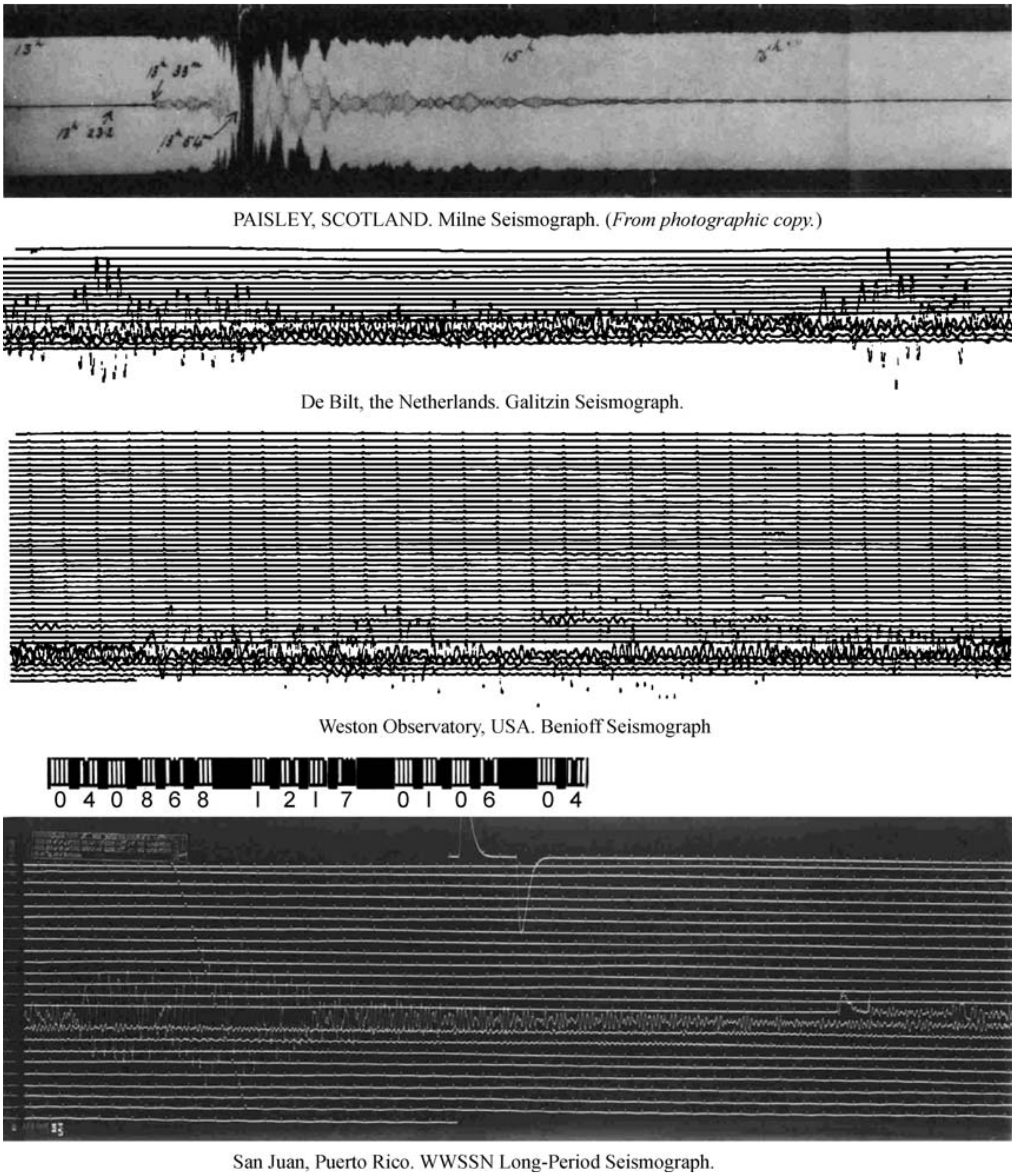


Earthquake Monitoring and Early Warning Systems, Figure 5
Some sample analog seismograms recorded on smoked paper

smaller, nearby earthquakes, while long-period instruments were designed to recover the motions of distant, larger earthquakes ("teleseisms"). Additionally, strong-motion accelerometers, generally recording directly onto 70 mm-wide film strips, were used to measure large motions from nearby earthquakes. In today's much more capable digital instrumentation, two major types of instruments are deployed: (1) "broadband" seismometers, which replace and improve upon both short-period and long-period seismometers, and (2) strong-motion accelerometers for high-amplitude, high-frequency, seismic waves from local earthquakes, which often drive broadband seismometers off scale. While rare examples of the old analog instruments are still in use, the vast majority of instruments presently operating are digital.

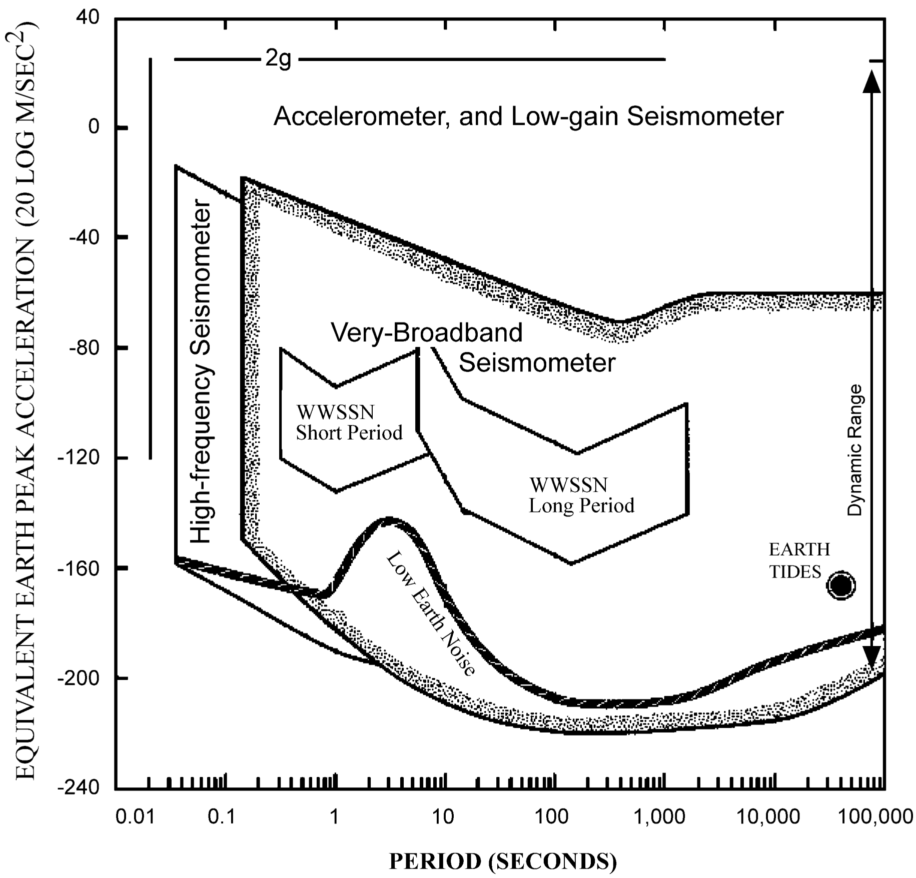
In addition to having large variations in amplitudes and frequencies, seismic waves from earthquakes also attenuate rapidly with distance, that is, they lose energy as they travel, particularly at higher frequencies. We must consider these effects in order to monitor seismic waves effectively.

In 1935, C.F. Richter introduced the concept of *magnitude* to classify local earthquakes by their "size", effectively the amount of energy radiated at the actual rup-



Earthquake Monitoring and Early Warning Systems, Figure 6
Some sample analog seismograms recorded on photographic paper or film

ture surface within the Earth. See the entry by Bormann and Saul ► Earthquake Magnitude for a discussion of the various magnitude scales in use. While every effort is made to make these different scales overlap cleanly, each has strengths and weaknesses that make one or another preferable in a given situation. Probably the most general and robust of these methods is called a "moment magnitude", symbolized as M_W. Existing instruments and environments are such that the smallest natural earthquakes we routinely observe close by are about magnitude $= 1$. The largest earthquake so far recorded by instrumentals is the $M_W = 9.5$ Chilean earthquake in 1960. In 1941, B.



Earthquake Monitoring and Early Warning Systems, Figure 7
Relative dynamic range of some common seismometers for global earthquake monitoring (modified from Fig. 1 in [54]). The *Y*-axis is marked in decibel (dB) where dB = 20 log(A/A_0); *A* is the signal amplitude, and A_0 is the reference signal amplitude

Gutenberg and C.F. Richter discovered that over large geographic regions the rate of earthquake occurrence is empirically related to their magnitudes by:

$$\log N = a - bM \qquad (1)$$

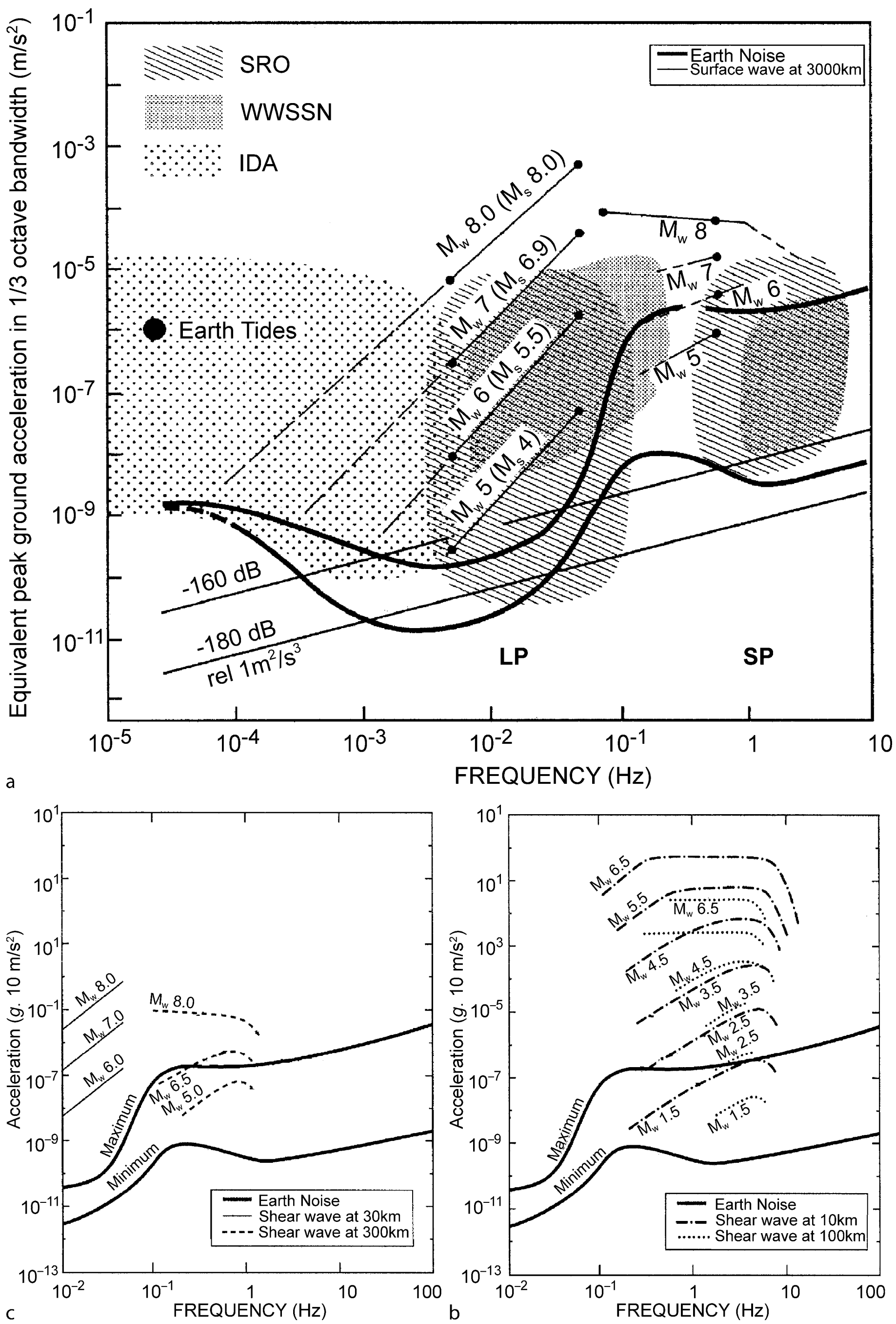
where N is the number of earthquakes of magnitude M or greater, and a and b are numerical constants. It turns out that b is usually about 1, which implies that $M = 6$ earthquakes are about ten times more frequent than $M = 7$ earthquakes. Engdahl and Villasenor [24] show that there has been an *average* of about 15 major ($M \geqslant 7$) earthquakes per year over the past 100 years, and about 150 large ($M \geq 6$) earthquakes per year during this same time interval. Strong ground motions (above 0.1 g in acceleration) over sizeable areas are generated by $M \geq 6$ earthquakes; these are potentially damaging levels of ground shaking.

Earthquakes are classified by magnitude (M) as *major* if $M \geq 7$, as *moderate* to *large* if M ranges from 5 to 7, as *small* if M ranges from 3 to 5, as *micro* if $M < 3$, and as *nano* if $M < 0$. An earthquake with $M \geq 7\,3/4$ is often called *great*, and if $M \geq 9$, *mega*.

Earthquake Monitoring in the Digital Era

Figure 8 shows the expected amplitudes of seismic waves by earthquake magnitude. The top frame is a plot of the equivalent peak ground acceleration versus frequency. The two heavy curves denote the "minimum Earth noise", and the "maximum Earth noise" (i. e., for seismographic station located in the continental interior versus near the coast).

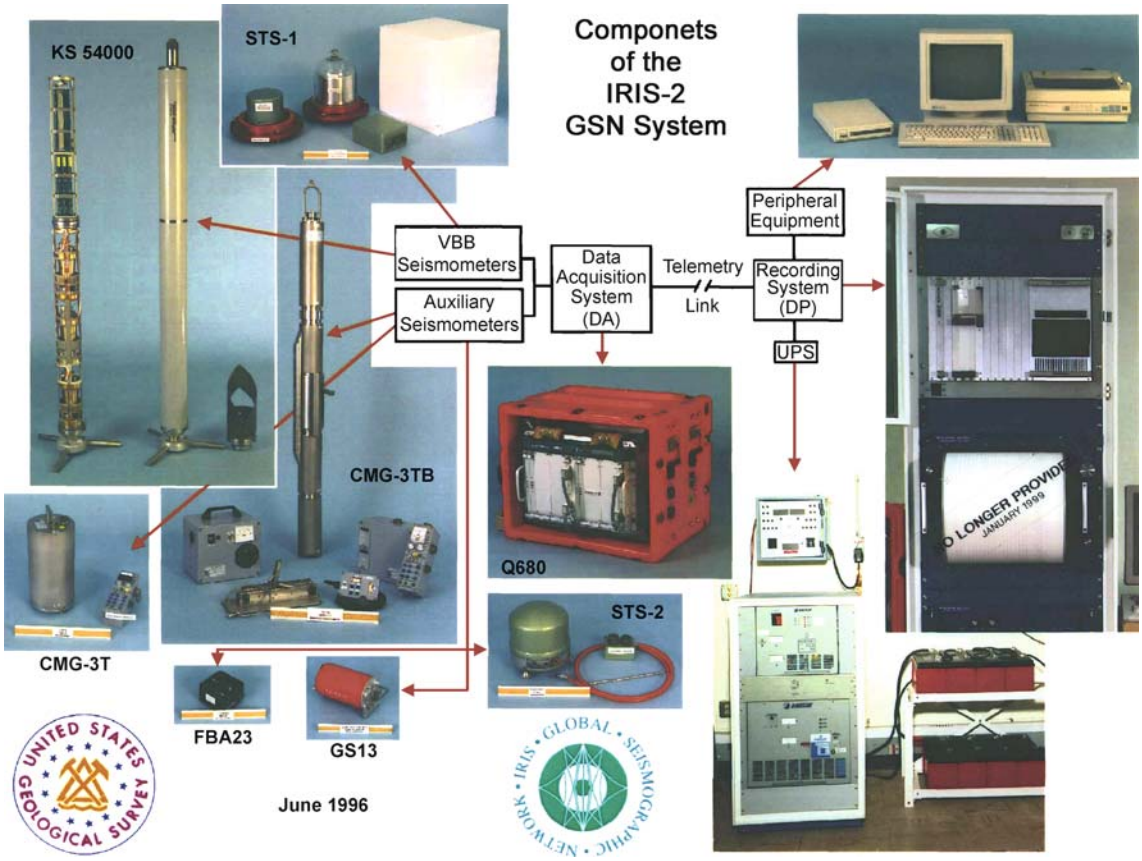
The two domains of the WWSSN equipment, short-period long-period seismometers are shown as gray shading. The domains for two other instruments, SRO (Seismic Research Observatories Seismograph) and IDA (International Deployment of Accelerometers), are also shown; these were the early models of the current instruments



Earthquake Monitoring and Early Warning Systems, Figure 8

Expected amplitudes of seismic waves by earthquake magnitude. See text for explanations

now in operation in the *Global Seismographic Network* (GSN). The bottom two frames indicate expected amplitudes of seismic waves from earthquakes of a range of magnitudes (we use the moment magnitude, M_W). For simplicity, we consider two cases: (bottom left) global earthquakes recorded at a large distance with a seismographic network spaced at intervals of about 1000 km, and (bottom right) local earthquakes recorded at short dis-

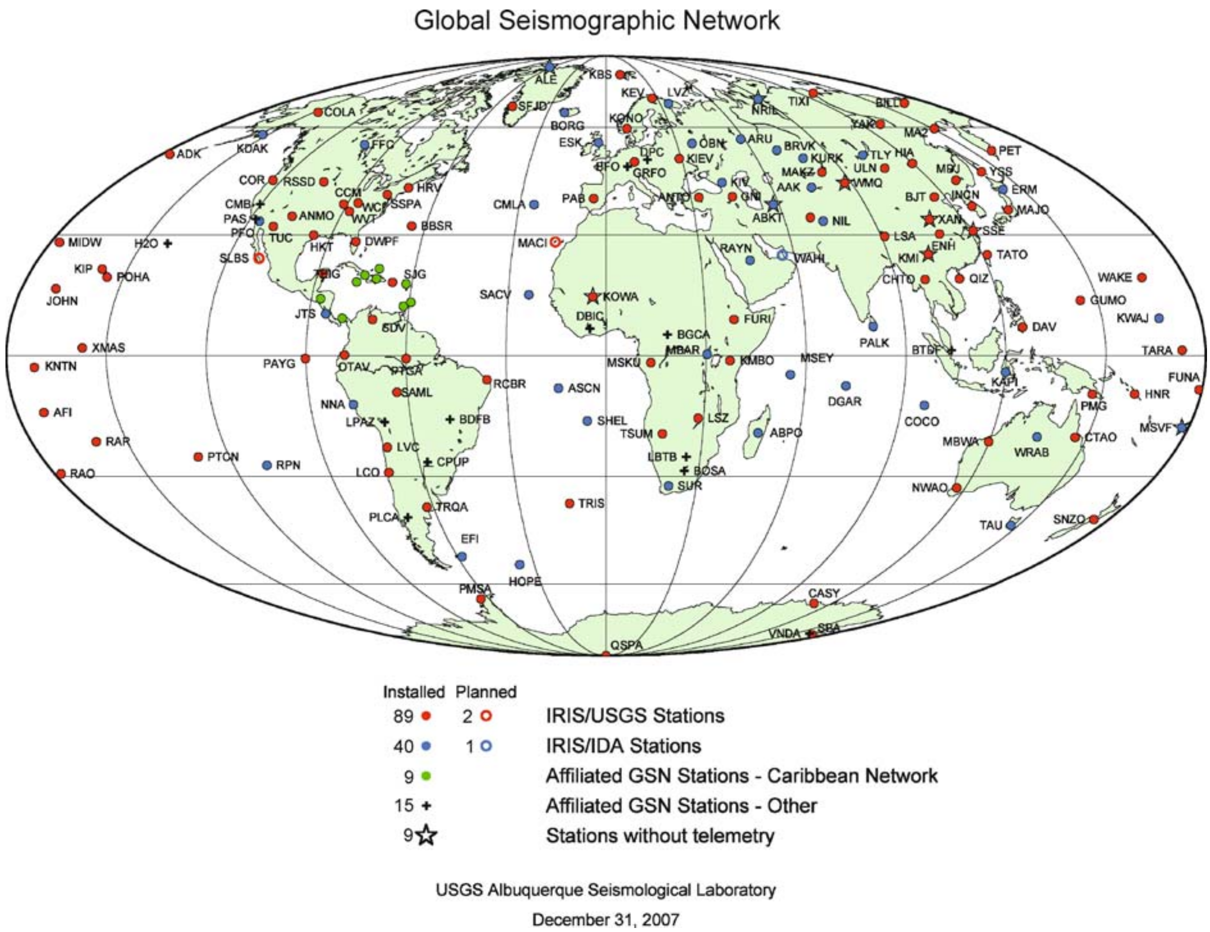


Earthquake Monitoring and Early Warning Systems, Figure 9
Components of the IRIS-2 GSN System: broadband seismometers, accelerometers and recording equipment

tances with a seismic array spaced at intervals of about 50 km. In the bottom left plot, the global-scale network, the expected amplitudes of *P*-wave and surface wave at 3000 km from the earthquake source are shown; for the bottom right plot, a local seismic array, the expected amplitudes of *S*-wave at 10 km and 100 km from the earthquake source are shown. Seismologists use this and similar figures in planning seismographic networks. Local noise surveys are usually conducted as well when designing specific seismographic networks.

With advances in digital technology, earthquake monitoring entered the digital era in the 1980s. Older analog equipment was gradually phased out as modern digital equipment replaced it [54]. The WWSSN was replaced by the *Global Seismographic Network* (GSN), a collaboration of several institutions under the IRIS consortium (http://www.iris.edu/). The goal of the GSN (http://www.iris.edu/about/GSN/index.htm) is "to deploy over 128 permanent seismic recording stations uniformly over the Earth's surface". The GSN project provides funding for two network operators: (1) the IRIS/ASL Network Operations Center, in Albuquerque, New Mexico (operated by the US Geological Survey), and (2) the IRIS/IDA Network Operations Center in La Jolla, California (operated by personnel from the Scripps Institution of Oceanography). Components of a modern IRIS GSN seismograph system, which include broadband seismometers, accelerometers, and recording equipment, are shown in Fig. 9.

Figure 10 shows the station map of the Global Seismographic Network as of 2007. IRIS GSN stations continuously record seismic data from very broad band seismometers at 20 samples per second (sps), and also include high-frequency (40 sps) and strong-motion (1 and 100 sps) sensors where scientifically warranted. It is the goal of the GSN project to provide real-time access to its data via Internet or satellite. Since 1991, the IRIS Data Management Center has been providing easy access to comprehensive seismic data from the GSN and elsewhere [1].



Earthquake Monitoring and Early Warning Systems, Figure 10
Station map of the Global Seismographic Network (GSN) as of 2007

Earthquake Monitoring: Regional and Local Networks

A major development in earthquake monitoring was the establishment of seismographic networks optimized to record the many *frequent* but *smaller* regional and local earthquakes occurring in many locations. To observe as many of these nearby earthquakes as possible, inexpensive seismographs with high magnifications and low dynamic-range telemetry are used to record the smallest earthquakes feasible with current technology and local background noise. As a result, the recorded amplitudes often overdrive the instruments for earthquakes with $M \gtrsim 3$ within about 50 km of such seismographs. This is not a serious defect, since the emphasis for these networks is to obtain as many first arrival times as possible, and to detect and to locate the maximum number of earthquakes. Because seismic waves from small earthquakes are quickly attenuated with increasing distance, it is also necessary to deploy many instruments at small station spacing (generally from a few to a few tens of kilometers), and to cover as large a territory as possible in order to record at least a few earthquakes every week. Since funding often is limited, these local and regional seismic networks are commonly optimized for the largest number of stations rather than for the highest quality data.

A Brief History

In the 1910s, the Carnegie Institution of Washington D.C. (CIW) was spending a great deal of money building the world's then largest telescope (100 inch) at Mount Wilson Observatory, southern California [38]. Since astronomers were concerned about earthquakes that might disturb their telescopes, Harry O. Wood was able to persuade CIW to support earthquake investigations, and as a result, a regional network of about a dozen Wood–Anderson seismographs was established in southern California in the 1920s. See Goodstein [38] for the early history leading to the establishment of the California Institute of Technology (Caltech) and its Seismological Laboratory. Astronomers played important roles in getting seismic monitoring established in various other regions of the world as well.

Regional networks using different types of seismographs were established in many countries about this time, such as in Japan, New Zealand, and the USSR and its allies. In the 1960s, high-gain, short-period, telemetered networks were developed to study microearthquakes. To support detailed studies of local earthquakes and especially for the purpose of earthquake prediction, over 100 microearthquake networks were established in various parts of the world by the end of the 1970s [74]. These microearthquake networks comprised from tens to hundreds of short-period seismometers, generally with their signals telemetered into central recording sites for processing and analysis. High magnification was achieved through electronic amplification, permitting recording of very small earthquakes (down to $M = 0$), though this came at the expense of saturated records for earthquakes of $M \gtrsim 3$ within about 50 km. Unfortunately, the hope of discovering some sort of earthquake precursor from the data obtained by these microearthquake networks did not work out. For a review of the earthquake prediction efforts, please read Kanamori [60].

Some Recent Advances

Because of recent advances in electronics, communications, and microcomputers, it is now possible to deploy sophisticated digital seismograph stations at global, national, regional, and local scales for *real-time* seismology [64]. Many such networks, including temporary portable networks, have been implemented in many countries. In particular, various real-time and near real-time seismic systems began operation in the 1990s: for example, in Mexico [25], California [32,47], and Taiwan [110]. The Real-Time Data (RTD) system operated by the Central Weather Bureau (CWB) of Taiwan is based on a network of telemetered digital accelerographs [102]; since 1995, this system has used pagers, e-mail, and other techniques to automatically and rapidly disseminate information about the hypocenter, magnitude, and shaking amplitude of felt earthquakes ($M \gtrsim 4$) in the Taiwan region. The disastrous Chi-Chi earthquake ($M_W = 7.6$) of 20 September 1999 caused 2,471 deaths and total economic losses of US$ 11.5 billion. For this earthquake sequence, the RTD system delivered accurate information to government officials 102 seconds after the origin time of the main shock (about 50 seconds for most aftershocks), and proved to be useful in the emergency response of the Taiwan government [37,131].

Recording Damaging Ground Shaking

Observing teleseisms on a global scale with station spacing of several hundreds of kilometers does not yield critical information about near-source strong ground shaking required for earthquake structural engineering purposes and seismic hazard reduction. Broadband seismometers, which are optimized to record earthquakes at great distances, do not perform well in the near-field of a major earthquake. For example, during the 1999 Chi–Chi earthquake the nearest broadband station in Taiwan (epicentral distance of about 20 km) was badly overdriven, recorded no useful data beyond the arrival time of the initial *P*-wave, and finally failed about one minute into the shock.

A regional seismic network with station spacing of a few tens of kilometers cannot do the job either: the station spacing is still too large and the records are typically overdriven for earthquakes of $M \gtrsim 3$ (any large earthquake would certainly overdrive these sensitive instruments in the entire network). In his account of early earthquake engineering, Housner [51] credited John R. Freeman, an eminent engineer, with persuading the then US Secretary of Commerce to authorize a strong-motion program, and, in 1930, the design of an accelerograph for engineering purposes. In a letter to R.R. Martel, Housner's professor at Caltech, Freeman wrote:

> *I stated that the data which had been given to structural engineers on acceleration and limits of motion in earthquakes as a basis for their designs were all based on guesswork, that there had never yet been a precise measurement of acceleration made. That of the five seismographs around San Francisco Bay which tried to record the earthquake of 1906 not one was able to tell the truth.*

Strong-motion recordings useful to engineers must be on-scale for damaging earthquakes and, in particular, from instruments located on or near built structures in densely urbanized environments, within about 25 km of the earthquake-rupture zone for sites on rock, or within about 100 km for sites on soft soils. Recordings of motions sufficient to cause damage at sites at greater distances are also of interest for earthquake engineering in areas likely to be affected by major subduction-zone earthquakes and in areas with exceptionally low attenuation rates [11]. In addition, densely-spaced networks of strong-motion recorders are needed to study the large variations in these motions over short distances [26,29].

Although several interesting accelerograms were recorded in southern California in the 1930s and 1940s, most seismologists did not pursue strong-motion monitoring until much later. The 1971 San Fernando earthquake emphatically demonstrated the need for more strong-motion data [9]. Two important programs emerged in the United States – the National Strong-Motion Program (http://nsmp.wr.usgs.gov/), and the California Strong Motion Instrumentation Program (http://docinet3.consrv.ca.gov/csmip/). However, the budgets for these programs were and continue to be small in comparison to other earthquake programs. High levels of funding for strong-motion monitoring, comparable to that of the GSN and the regional seismic networks, became available in Taiwan in the early 1990s, and in Japan in the mid-1990s. The Consortium of Organizations for Strong-Motion Observation Systems (http://www.cosmos-eq.org/) was established recently to promote the acquisition and application of strong-motion data.

Seismograms and Derived Products

Even before instruments were developed to record seismic waves from earthquakes, many scholars compiled catalogs of earthquake events noted in historical and other documents. Robert Mallet in 1852–1854 published the first extensive earthquake catalog of the world (1606 B.C.–A.D. 1842) totaling 6831 events [79]. Based on this compilation, Mallet prepared the first significant seismicity map of the Earth in 1858. Mallet's map is remarkable in that it correctly identifies the major earthquake zones of the Earth excepting for parts of the oceans. Although Mallet's earthquake catalog and similar compilations contain a wealth of information about earthquakes, they were made without the aid of instruments, and thus were subject to the biases of the observers as well as to population distributions. These non-instrumental earthquake catalogs also contain errors because the source materials were commonly incomplete and inconsistent regarding date, time, place names, and reported damage. Ambraseys et al. [8] discusses these difficulties for a regional case and Guidoboni [42] addresses the matter in general.

Today, seismograms are the fundamental data produced by earthquake monitoring. An analyst's first task is to find out when and where the earthquakes occurred, its size, and other characteristics. The accuracy of determining earthquake parameters, as well as the number of parameters used to characterize and earthquake, has progressed along with the availability of seismograms and computers, as well as advances in seismology. In the analog era, earthquake parameters were primarily the origin time, geographical location (epicenter), focal depth, and magnitude. A list of these parameters for earthquakes occurring over some time interval is called an *earthquake catalog*. A useful and common illustration of such results is a map showing the locations of earthquakes by magnitude (a seismicity map). Figure 11 is such a seismicity map for 1900–1999 as prepared by Engdahl and Villasenor [24]. The map shows that moderate and large earthquakes are concentrated in tectonic active areas while most areas of the Earth are aseismic.

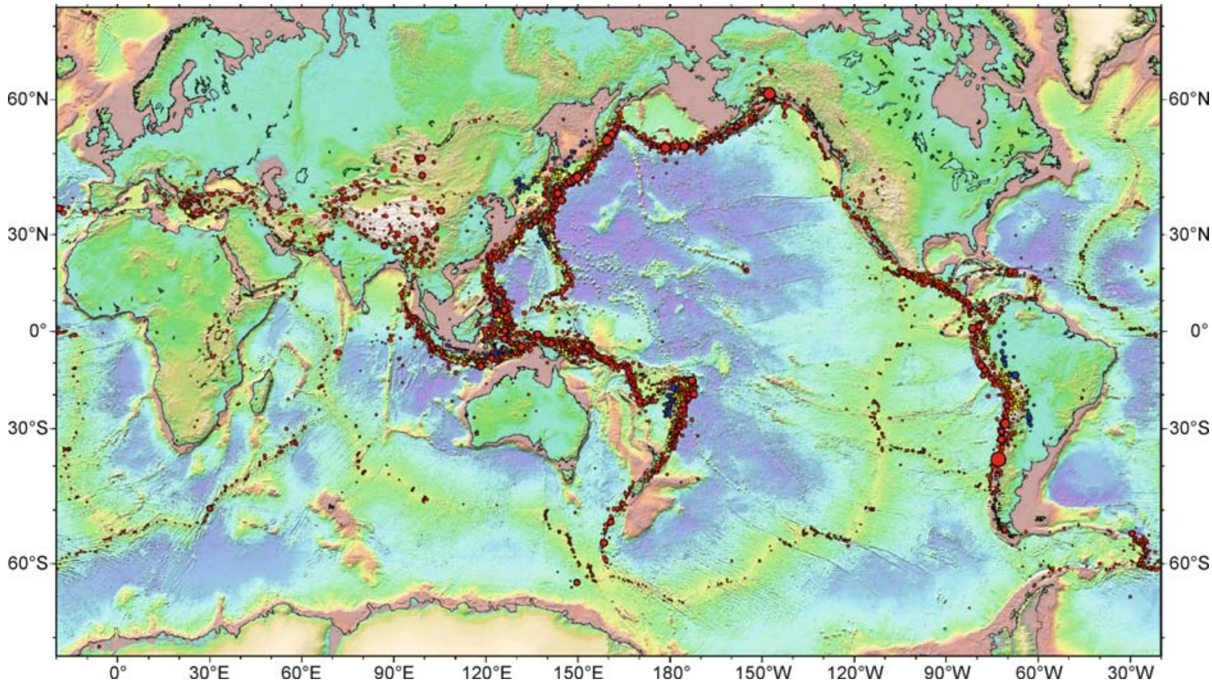
Earthquake Location

Several methods have been developed to locate earthquakes (i. e., determine origin time, latitude and longitude of the epicenter, and focal depth). Common to most of these methods is the use of arrivals times of initial *P*- and *S*-waves. In particular, Geiger [33] applied the Gauss–Newton method to solve for earthquake location, which is a nonlinear problem, by formulating it as an inverse problem. However, since Geiger's method is computational intensive, it was not practical to apply it for the routine determinations of earthquake hypocenters until the advance of modern computers in the early 1960s.

Before computers became widely available starting in the 1960s, earthquakes were usually located by a manual, graphical method. In any location method, we assume that an earthquake is a point source and its sole parameters are origin time (time of occurrence, t_o) and hypocenter position (x_o, y_o, z_o). If both *P*- and *S*-arrival times are available, one may use the time intervals between *P*- and *S*-waves at each station (*S*-*P* times) and estimates of seismic wave velocities in the Earth to obtain a rough estimate of the epicentral distance, *D*, from that station:

$$D = [V_P V_S/(V_P - V_S)](T_S - T_P) \qquad (2)$$

where V_P is the *P*-wave velocity, V_S the *S*-wave velocity, T_S the *S*-wave arrival time, and T_P the *P*-wave arrival time. For a typical crustal *P*-wave velocity of 6 km/s, and $V_P/V_S \approx 1.8$, the distance *D* in kilometers is about 7.5 times the *S*-*P* interval measured in seconds. If three or more epicentral distances are available, the epicenter may be placed at the intersection of circles with the stations as centers and the appropriate *D* as radii. The intersection



Earthquake Monitoring and Early Warning Systems, Figure 11
Seismicity of the Earth: 1900–1999 (see [24] for details)

will seldom be a point, and its areal extent gives a rough estimate of the uncertainty of the epicenter and hypocentral (focal) depth. In the early days, the focal depth was usually assumed or occasionally determined using a "depth phase" (generally, a ray that travels upward from the hypocenter and reflects back from the Earth's surface, then arcs through the Earth to reach a distant seismograph).

Although Geiger [33] presented a method for determining the origin time and epicenter, the method can be extended easily to include focal depth. To locate an earthquake using a set of arrival times, τ_k, from stations at positions (x_k, y_k, z_k), $k = 1, 2, \ldots, m$, we must assume a model of seismic velocities from which theoretical travel times, T_k for a trial hypocenter at (x^*, y^*, z^*) to the stations can be computed. Let us consider a given trial origin time and hypocenter as the trial vector $\boldsymbol{\chi}^*$ in a four-dimensional Euclidean space:

$$\boldsymbol{\chi}^* = (t^*, x^*, y^*, z^*)^{\mathrm{T}} \tag{3}$$

where the superscript T ($^{\mathrm{T}}$) denotes the vector transpose. Theoretical arrival time, t_k, from $\boldsymbol{\chi}^*$ to the k-th station is the theoretical travel time, T_k, plus the trial origin time, t^*. We now define the arrival time residual at the k-th station, r_k, as the difference between the observed and the theoretical arrival times. We may consider this set of station residuals as a vector in an m-dimensional Euclidean space and write:

$$\mathbf{r} = (r_1(\boldsymbol{\chi}^*), \mathrm{r}_2(\boldsymbol{\chi}^*), \ldots, \mathrm{r}_m(\boldsymbol{\chi}^*))^{\mathrm{T}} \,. \tag{4}$$

We now apply the least squares method to obtain a set of linear equations solving for an adjustment vector, $\boldsymbol{\delta\chi}$:

$$\mathbf{A}^{\mathrm{T}}\mathbf{A}\boldsymbol{\delta\chi} = -\mathbf{A}^{\mathrm{T}}\mathbf{r} \,, \tag{5}$$

where A is the Jacobian matrix consisting of partial derivatives of travel time with respect to t, x, y, and z. A detailed derivation of the Geiger method is given by Lee and Stewart (see, pp 132–134 in [74]). There are many practical difficulties in implementing Geiger's method for locating earthquakes, as discussed by Lee and Stewart (see, pp 134–139 in [74]). Although standard errors for these earthquake locations can be computed, they are often not meaningful because errors in the measurement of arrival times usually do not obey a Gaussian probability distribution. In recent years, many authors applied various nonlinear methods to locate earthquakes; a review of these methods is given by Lomax et al. ► Earthquake Location, Direct, Global-Search Methods.

Earthquake Magnitude

After an earthquake is located, the next question is: how big was it? The Richter magnitude scale was originally devised to measure the "size" of an earthquake in southern California. Richter [96] defined the local (earthquake) magnitude, M_L, of an earthquake observed at any particular station to be:

$$M_L = \log A - \log A_0(\Delta) \tag{6}$$

where A is the maximum amplitude in millimeters as recorded by a Wood–Anderson seismograph for an earthquake at epicentral distance of Δ km. The correction factor, $\log A_0(\Delta)$, is the maximum amplitude at Δ km for a "standard" earthquake. Thus, three arbitrary choices enter into the definition of local magnitude: (1) the use of the Wood–Anderson seismographs, (2) the use of the common logarithm (i. e., the logarithm to the base 10), and (3) the selection of the standard earthquake, whose amplitudes as a function of distance Δ are represented by $A_0(\Delta)$.

In the 1940s, B. Gutenberg and C.F. Richter extended the local magnitude scale to include more distant earthquakes. Gutenberg [43] defined the surface-wave magnitude, M_S, as

$$M_S = \log(A/T) - \log A_0(\Delta^\circ) \tag{7}$$

where A is the maximum combined horizontal ground displacement in micrometers (μm) for surface waves with a period of 20 s, and $-\log A_0$ is tabulated as a function of epicentral distance Δ in degrees, in a similar manner to that for the local magnitude's A_0 (Δ). Specifically, surface-wave magnitude is calculated from

$$M_S = \log A + 1.656 \log \Delta + 1.818 \tag{8}$$

using the prominent 20 s period surface waves commonly observed on the two horizontal-component seismograms from earthquakes of shallow focal depth.

Both magnitude scales were derived empirically and have scale-saturation problems, e. g., for very large earthquakes above a certain size the computed magnitudes of a particular type are all about the same After the pioneering work of Charles F. Richter and Beno Gutenberg, numerous authors have developed alternative magnitude scales, as reviewed recently by Utsu [116] and by Bormann and Saul ► Earthquake Magnitude. A current magnitude scale widely accepted as "best" (as having the least saturation problem and being a close match to an earthquake's total release of stress and strain) is the "moment magnitude", M_W, computed from an earthquake's "moment tensor".

Quantification of the Earthquake Source

As pointed out by Kanamori [59], it is not a simple matter to find a single measure of the "size" of an earthquake, simply because earthquakes result from complex physical processes. The elastic rebound theory of Harry F. Reid suggests that earthquakes originate from spontaneous slippage on active faults after a long period of elastic strain accumulation [94]. Faults may be considered the slip surfaces across which discontinuous displacement occurs in the Earth, while the faulting process may be modeled mathematically as a shear dislocation in an elastic medium (see [100], for a review). A shear dislocation (or slip) is equivalent to a double-couple body force [15,81]. The scaling parameter of each component couple of a double-couple body force is its *moment*. Using the equivalence between slip and body forces, Aki [2] introduced the *seismic moment*, M_0, as:

$$M_0 = \mu \int D(A)\mathrm{d}A = \mu s A \tag{9}$$

where μ is the shear modulus of the medium, A is the area of the slipped surface or source area, and s is the slip $D(A)$ averaged over the area A. If an earthquake produces surface faulting, we may estimate its rupture length, L, and its average slip, s, from measurement of that faulting. The area A may be approximated by Lh, where h is the focal depth (it is often, but not always, found that the hypocenter is near the bottom of the rupture surface). A reasonable estimate for μ is 3×10^{11} dynes/cm^2. With these quantities, we can estimate the seismic moment from Eq. (9).

Seismic moment also can be estimated independently from seismograms. From dislocation theory, the seismic moment can be related to the far-field seismic displacement recorded by seismographs. For example, Hanks and Wyss [46] showed that

$$M_0 = (\Omega_0/\psi_{\theta\phi})4\pi\rho R v^3 \tag{10}$$

where Ω_0 is the long-period limit of the displacement spectrum of either P or S waves, $\psi_{\theta\phi}$ is a function accounting for the body-wave radiation pattern, ρ is the density of the medium, R is a function accounting for the geometric spreading of body waves, and v is the body-wave velocity. Similarly, seismic moment can be determined from surface waves or coda waves [2,3].

In 1977, Hiroo Kanamori recognized that a new magnitude scale can be developed using seismic moment (M_0) by comparing the earthquake energy and seismic moment relation

$$E_S = (\Delta\sigma/2\mu)M_0\,, \tag{11}$$

where $\Delta\sigma$ is the stress drop and μ is the shear modulus, with the surface-wave magnitude and energy relation [45]

$$\log E_S = 1.5M_S + 11.8\,, \tag{12}$$

where E_S and M_0 are expressed in ergs and dyne-cm, respectively. The average value of $(\Delta\sigma/2\mu)$ is approximately equal to 1.0×10^{-4}. If we use this value in Eq. (11), we obtain

$$\log M_0 = 1.5M_S + 16.1\,. \tag{13}$$

It is known that M_S values saturate for great earthquakes (M_0 about 10^{29} dyne − cm or more) and, therefore, that Eqs. (12) and (13) do not hold for such great earthquakes. If a new moment-magnitude scale using the notation M_W is defined by

$$\log M_0 = 1.5M_W + 16.1 \tag{14}$$

then M_W is equivalent to M_S below saturation and provides a reasonable estimate for great earthquakes without the saturation problem [58]. The subscript letter *W* stands for the work at an earthquake fault, but soon M_W became known as the *moment magnitude*. Determining earthquake magnitude using seismic moment is clearly a better approach because it has a physical basis.

The concept of seismic moment led to the development of moment tensor solutions for quantifying the earthquake source, including its focal mechanism [35,36]; the seismic moment is just the scalar value of the moment tensor. Since the 1980s, Centroid-Moment-Tensor (CMT) solutions have been produced routinely for events with moment magnitudes (M_W) greater than about 5.5. The CMT methodology is described by Dziewonski et al. [22] and Dziewonski and Woodhouse [20]; a comprehensive review is given in Dziewonski and Woodhouse [21]. These CMT solutions are published yearly in the journal *Physics of the Earth and Planetary Interiors*, and the entire database is accessible online. This useful service is now performed by the Global CMT Project (http://www.globalcmt.org/), and more than 25,000 moment tensors have been determined for large earthquakes from 1976 to 2007. In the most recent decade, Quick CMT solutions [23] determined in near-real time have been added and are distributed widely via e-mail (http://www.seismology.harvard.edu/projects/CMT/QuickCMTs/).

Limitations of Earthquake Catalogs

In addition to international efforts to catalog earthquakes on a global scale, observatories and government agencies issue more-detailed earthquake catalogs at local, regional, and national scales. However, earthquake catalogs from local to global scales vary greatly in spatial and temporal coverage and in quality, with respect to completeness and accuracy, because of the ongoing evolution of instrumentation, data processing procedures, and agency staff. An earthquake catalog, to be used for research, should have at least the following source parameters: origin time, epicenter (latitude and longitude), focal depth, and magnitude.

The International Seismological Summary and its predecessors provided compilations of arrival times and locations of earthquakes determined manually from about 1900 to 1963. Despite their limitations (notably the lack of magnitude estimates), these materials remain valuable. The first global earthquake catalog that contains both locations and magnitudes was published by Gutenberg and Richter in 1949, and was followed by a second edition in 1954 [44]. This catalog contains over 4,000 earthquakes from 1904 to 1951. Unfortunately, its temporal and spatial coverage is uneven as a result of rapid changes in seismic instrumentation, and of the interference of both World Wars. Nevertheless, the procedures used for earthquake location and magnitude estimation were the same throughout, using the arrival-time and amplitude data available to Gutenberg and Richter during the 1940s and early 1950s.

Since 1964, the International Seismological Centre has performed systematic cataloging of earthquakes worldwide by using computers and more modern seismograph networks. The spatial coverage of this catalog is not complete for some areas of the Earth (especially the oceans) because of the paucity of seismographic stations in such areas. By plotting the cumulative numbers of earthquakes above a certain magnitude versus magnitude, and using Eq. (1), the lower limit of completeness of an earthquake catalog may be estimated – it is the magnitude below which the data deviate below a linear fit to Eq. (1).

A *Centennial Earthquake Catalog* covering ISS- and ISC-reported global earthquakes from 1900–1999 was generated using an improved Earth model that takes into account regional variations in seismic wave velocities in the Earth's crust and upper mantle [24,118]. Engdahl and Villasenor [24] also compiled existing magnitude data from various authors and suggested preferred values. However, these "preferred magnitudes" were not determined by the same procedures. At present, the Global CMT Project (http://www.globalcmt.org/) provides the most complete online source parameters for global earthquakes (with $M_W > 5.5$), including Centroid-Moment-Tensor solutions. Although the CMT catalog starts in 1976, the improved global coverage of modern broadband digital seismographs began only in about 1990.

In summary, earthquake catalogs have been used extensively for earthquake prediction research and seismic hazard assessment since the first such catalog was produced. Reservations have been expressed about the reliability of the results and interpretations from these studies because the catalogs cover too little time and have limitations in completeness and accuracy (both random and systematic). Nevertheless, advances have been made in using earthquake catalogs to (1) study the nature of seismicity (e. g., ► Seismicity, Critical States of: From Models to Practical Seismic Hazard Estimates Space), (2) investigate earthquake statistics (e. g., ► Earthquake Occurrence and Mechanisms, Stochastic Models for), (3) forecast earthquakes (e. g., ► Earthquake Forecasting and Verification), (4) predict earthquakes (e. g., ► Geo-complexity and Earthquake Prediction), (5) assess seismic hazards and risk, and so forth.

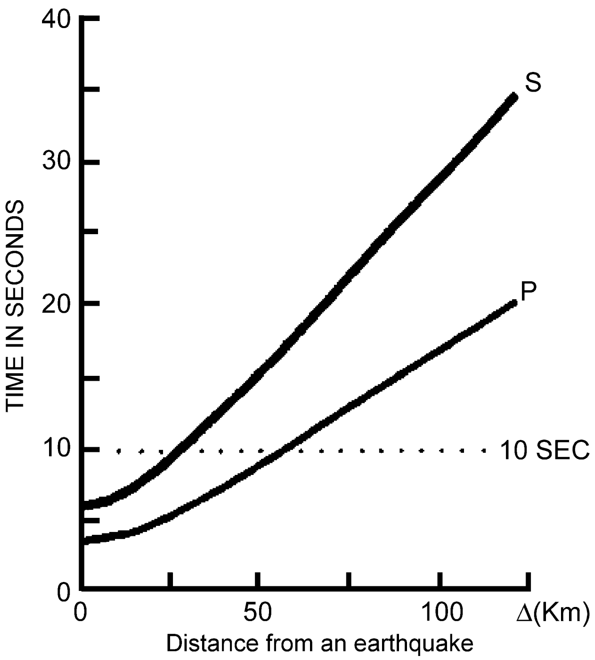
Earthquake Early Warning (EEW) Systems

With increasing urbanization worldwide, earthquake hazards pose ever greater threats to lives, property, and livelihoods in populated areas near major active faults on land or near offshore subduction zones. Earthquake early-warning systems can be useful tools for reducing the impact of earthquakes, provided that cities are favorably located with respect to earthquake sources and their citizens are properly trained to respond to the warning messages. Recent reviews of earthquake early warning systems may be found in Lee and Espinosa-Aranda [73], Kanamori [61], and Allen [6], as well as a monograph on the subject by Gasparini et al. [31].

Under favorable conditions, an EEW system can forewarn an urban area of impending strong shaking with lead times that range from a few seconds to a few tens of seconds. A lead time is the time interval between issuing a warning and the arrival of the *S*-waves, which are the most destructive seismic waves. Even a few seconds of advanced warning is useful for pre-programmed emergency measures at various critical facilities, such as the deceleration of rapid-transit vehicles and high-speed trains, the orderly shutoff of gas pipelines, the controlled shutdown of some high-technological manufacturing operations, the safe-guarding of computer facilities (e. g., disk-head parking), and bringing elevators to a stop at the nearest floor.

Physical Basis and Limitations of EEW Systems

The physical basis for earthquake early warning is simple: damaging strong ground shaking is caused primarily by shear (*S*) and subsequent surface waves, both of which travel more slowly that the primary (*P*) waves, and seismic waves travel much more slowly than electromagnetic signals transmitted by telephone or radio. However, certain physical limitations must be considered, as shown by Fig. 12.



Earthquake Monitoring and Early Warning Systems, Figure 12
Travel time of *P*-waves and of *S*-waves versus distance for a typical earthquake

Figure 12 is a plot of the travel time for the *P*-wave and *S*-wave as a function of distance from an earthquake. We make the following assumptions about a typical destructive earthquake: (1) focal depth at ~ 20 km, (2) *P*-wave velocity ~ 8 km/s, and (3) *S*-wave velocity ~ 4.5 km/s. If an earthquake is located 100 km from a city, the *P*-wave arrives at the city after about 13 s, and the *S*-waves in about 22 s (Fig. 12). If we deploy a dense seismic network near the earthquake source area (capable of locating and determining the size of the event in about 10 s), we will have about 3 s to issue the warning before the *P*-wave arrives, and about 12 s before the more destructive *S*-waves and surface waves arrive at the city. We have assumed that it takes negligible time to send a signal from the seismic network to the city via electromagnetic waves, which travel at about one-third the velocity of light or faster (between about 100,000 and 300,000 km/s depending on the method of transmission).

From Fig. 12 it is clear that this strategy may work for earthquakes located at least about 60 km from the urban area. For earthquakes at shorter distances (~ 20 to

~ 60 km), we must reduce the time needed to detect the event and issue a warning to well under 10 s. This requirement implies that we must deploy a very dense seismic network very close to the fault and estimate the necessary parameters very fast. However, such dense networks are not economical to deploy using existing seismic instruments.

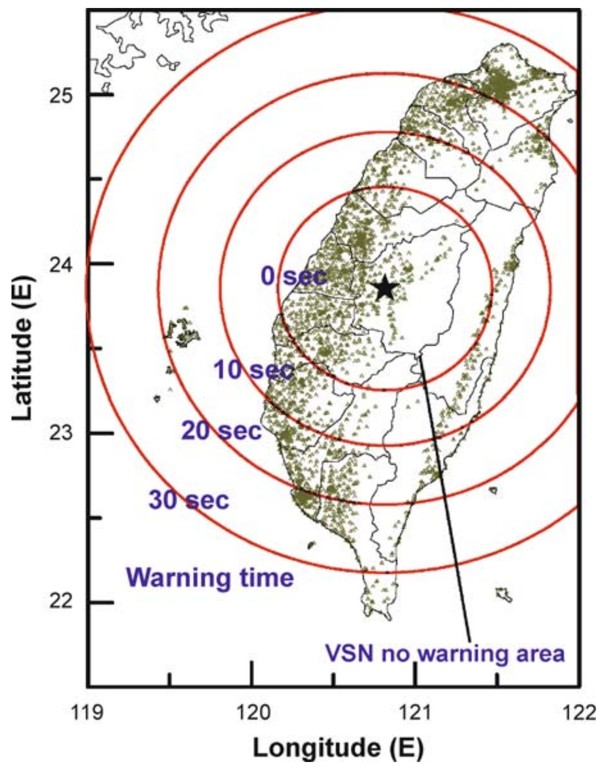
For earthquakes within 20 km of a city, there is little one can do other than installing motion-sensitive automatic shut-off devices at critical facilities (natural gas, for example) and hope that they are either very quick when responding to *S*-waves or are triggered by the onset of the *P*-wave. Normally an earthquake rupture more than ~ 100 km from an urban area does not commonly pose a large threat (seismic waves would be attenuated and spread out farther). There are exceptions caused either by unusual local site conditions, such as Mexico City, or by earthquakes with large rupture zones which therefore radiate efficiently to greater distances.

Design Considerations for EEW Systems

In the above discussion, we have assumed that one implements an earthquake early warning system with a traditional seismic network. Such EEW systems have limitation as illustrated by Fig. 13, which shows the expected early warning times for a repeat of the 1999 Chi–Chi earthquake. However, Nakamura and his colleagues have been successful in applying a single-station approach [84,99], where seismic signals are recorded and processed locally by the seismograph and an earthquake warning is issued whenever ground motions there exceed some trigger threshold. We will next discuss these two basic approaches, regional versus on-site in designing an earthquake early warning system.

Earthquake early warning capability can be implemented through a rapid reporting system (RRS) from a traditional network, assuming real-time telemetry into the network's central laboratory. This type of system provides, to populated areas and other sensitive locations, primary event information (hypocenter, magnitude, ground shaking intensities, and potential damage) about one minute after the earthquake begins. The RRS transmits this critical information electronically to emergency response agencies and other interested organizations and to individuals. Each recipient can then take action (some of which may be pre-programmed) shortly after the earthquake begins. Response measures can include the timely dispatch of rescue equipment and emergency supplies to the likely areas of damage.

California's ShakeMap [119,120], Taiwan's CWB, and Japan's JMA systems are typical examples of RSS. In



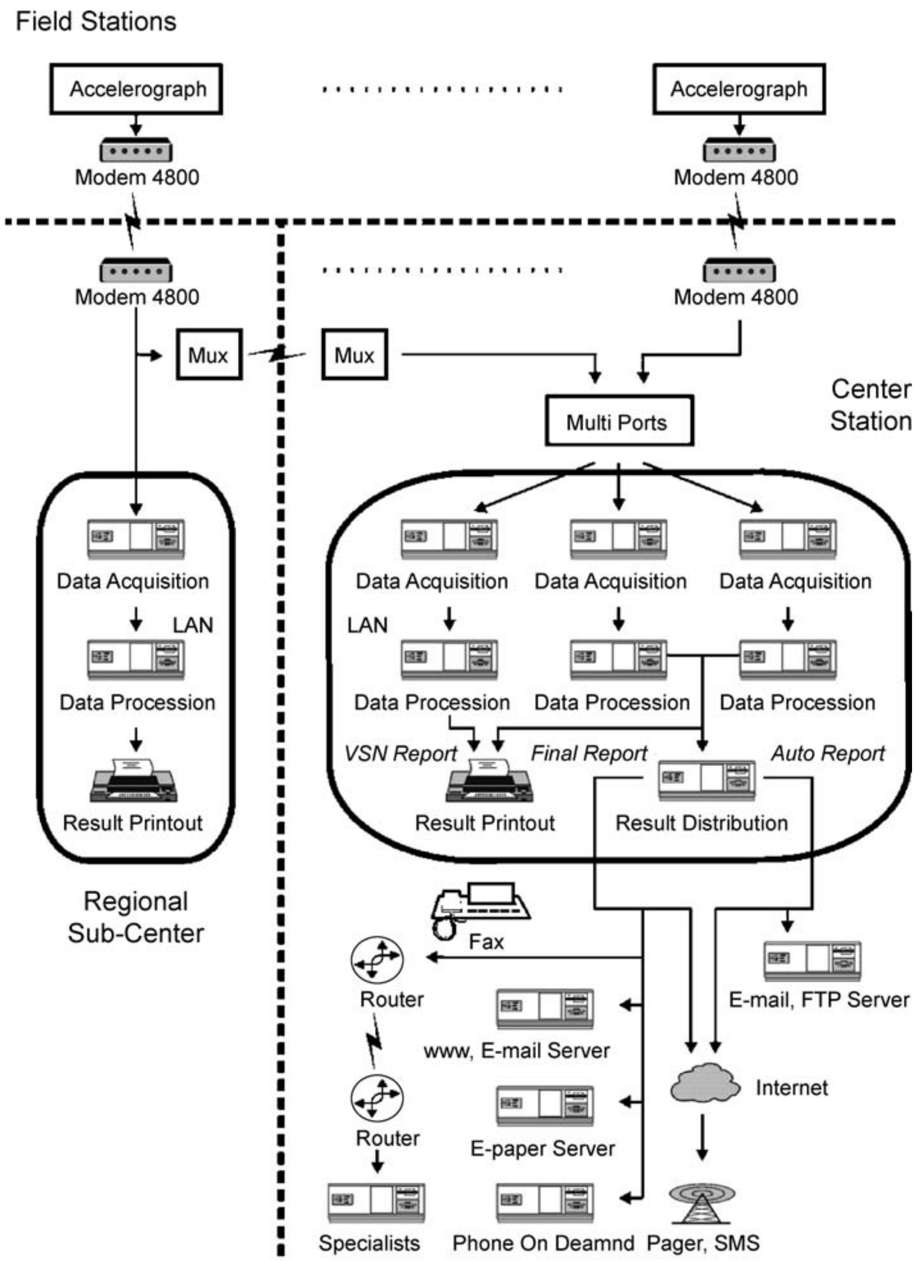
Earthquake Monitoring and Early Warning Systems, Figure 13
Expected EWS early warning times (indicated by *circles*) in Taiwan with respect to the occurrence of an event similar to the Chi–Chi earthquake of 20 September 1999. *Triangles* are locations of elementary schools, which can be regarded as a good indicator for the population density of Taiwan

the case of the Taiwan RRS, the CWB has, since 1995, provided intensity maps, hypocenters, and magnitudes within one minute of the occurrence of $M > 4$ earthquakes [110,128]. This system's reliability, documented by electronic messages to government agencies and scientists, has been close to perfect, particularly for large, damaging earthquakes. Figure 14 shows a block diagram of the Taiwan RRS, and details may be found in [128].

Using a set of empirical relationships derived from the large data set collected during the 1999 Chi–Chi earthquake, CWB now releases, within a few minutes of an event, the estimated distributions of PGA and PGV, refined magnitudes, and damage estimates [129]. This near-real-time damage assessment is useful for rapid post-disaster emergency response and rescue missions.

Regional Warning Versus Onsite Warning

Two approaches have been adopted for earthquake early warning systems: (1) regional warning, and (2) on-site



Earthquake Monitoring and Early Warning Systems, Figure 14
A block diagram showing the hardware of the Taiwan Earthquake Rapid Reporting System

warning. The first approach relies on traditional seismological methods in which data from a seismic network are used to locate an earthquake, determine the magnitude, and estimate the ground motion in the region involved. In the second approach, the initial ground motions (mainly *P* wave) observed at a site are used to predict the ensuing ground motions (mainly *S* and surface waves) at the same site.

The regional approach is more comprehensive, but takes a longer time to issue an earthquake warning. An advantage of this approach is that estimates of the timing of expected strong motions throughout the affected region can be predicted more reliably. The early warning system in Taiwan is a typical example and it uses a regional warning system called virtual sub-network approach (VSN) that requires an average of 22 s to determine earthquake parameters with magnitude uncertainties of ± 0.25. It provides a warning for areas beyond about 70 km from the epicenter (Fig. 13). This system has been in operation since 2002 with almost no false alarms [129]. With the advancement of new methodology and more dense seismic networks, regional systems are beginning to be able to provide early warnings to areas closer to the earthquake epicenter.

The regional approach has also been used in other areas. The method used in Mexico [25] is slightly different from the traditional seismological method. It is a special case of EEW system due to the relatively large distance (about 300 km in this case) between the earthquake source region (west coast of Central America) and the warning site (Mexico City). However, the warning is conceptually "regional".

In Japan, various EEW techniques have been developed and deployed by the National Research Institute for Earth Science and Disaster Prevention (NIED) and Japan Meteorological Agency (JMA) since 2000 [49,57,89], ► Tsunami Forecasting and Warning. In particular, JMA has started sending early warning messages to potential users responsible for emergency responses [50]. The potential users include railway systems, construction companies, and others; and they are familiar with the implications of early warning messages, as well as the technical limitations of EEW [57].

Some Recent EEW Advances

Allen and Kanamori [7] proposed the Earthquake Alarm System (ElarmS) to issue an earthquake warning based on information determined from the P-wave arrival only. Kanamori [61] extended the method of Nakamura [84] and Allen and Kanamori [7] to determine a period parameter, τ_c, from the initial 3 s of the P wave. τ_c is defined as

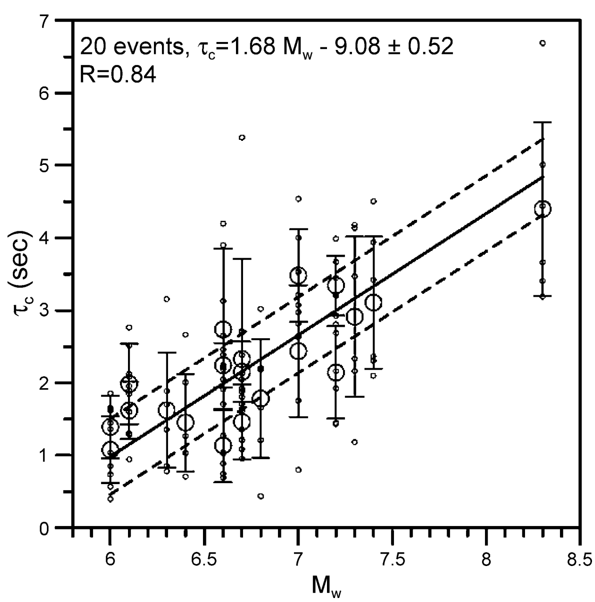
$$\tau_c = 2\pi/\sqrt{r} \tag{15}$$

where

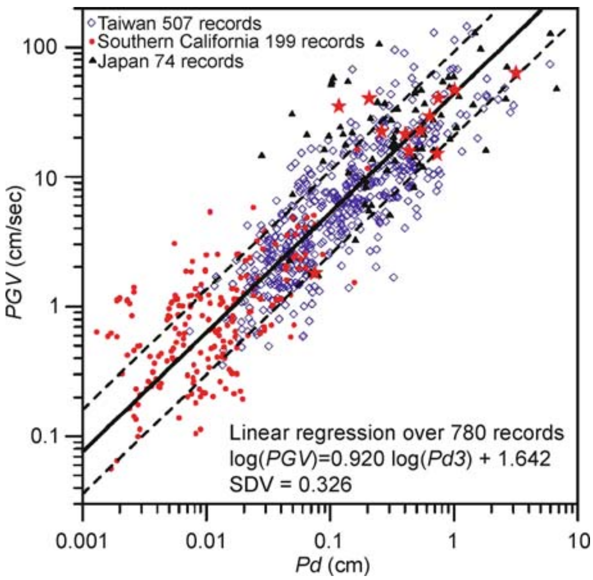
$$r = \frac{\int_0^{\tau_0} \dot{u}^2(t)\,dt}{\int_0^{\tau_0} u^2(t)\,dt} \tag{16}$$

$u(t)$ is the ground-motion displacement; τ_0 is the duration of record used (usually 3 s), and τ_c, which represents the size of an earthquake, can be computed from the incoming data sequentially.

The τ_c method was used for earthquake early warning in southern California, Taiwan, and Japan by Wu and Kanamori [124,125,126] and Wu et al. [130]. At a given site, the magnitude of an event is estimated from τ_c and the peak ground-motion velocity (PGV) from P_d (the peak amplitude of displacement in the first 3 s after the arrival of the P wave). The incoming 3-component signals are recursively converted to ground acceleration, velocity and displacement. The displacements are recursively filtered using an accusal Butterworth high-pass filter with a cut-off frequency of 0.075 Hz, and a P-wave threshold trigger is constantly monitored. When a trigger occurs, τ_c and P_d are computed. The relationships between τ_c and magnitude (M), and P_d and peak ground velocity (PGV) for southern California, Taiwan, and Japan were investigated. Figure 15 shows a good correlation between τ_c and M_W from the K-NET records in Japan, and Fig. 16 shows the P_d versus PGV plot for southern California, Taiwan, and Japan. These relationships may be used to detect the occurrence of a large earthquake and provide onsite warning in the area immediately around the station where the onset of strong ground motion is expected within a few seconds after the arrival of the P-wave. When the station density is high, the onsite warning methods may be applied to data from multiple stations to increase the robustness of an onsite early warning, and to complement the regional warning approach. In an ideal situation, such warnings would be available within 10 s of the origin time of a large earthquake whose subsequent ground motion may last for tens of seconds.



Earthquake Monitoring and Early Warning Systems, Figure 15
τ_c estimates from 20 events using the nearest 6 stations of the K-NET. *Small open circles* show single-record results, and large circles show event-average values with one standard deviation bars. *Solid line* shows the least squares fit to the event-average values, and the *two dashed lines* show the range of one standard deviation



Earthquake Monitoring and Early Warning Systems, Figure 16
Relationship between peak initial displacement amplitude (P_d) measurements and peak ground velocity (*PGV*) for the records with epicentral distances less than 30 km from the epicenter in Southern California (*red solid circles*), Taiwan (*blue diamonds*) and Japan (*black solid triangles*). *Solid line* shows the least squares fit and the two *dashed lines* show the range of one standard deviation

Wu and Zhao [127] investigated the attenuation of P_d with the hypocentral distance R in southern California as a function of magnitude M, and obtained the following relationships:

$$M_{P_d} = 4.748 + 1.371 \times \log(P_d) + 1.883 \times \log(R) \quad (17)$$

and

$$\log(P_d) = -3.463 + 0.729 \times M - 1.374 \times \log(R)\,. \quad (18)$$

For the regional warning approach, when an earthquake location is determined by the P-wave arrival times at stations close to the epicenter, this relationship can be used to estimate the earthquake magnitude. Their result shows that for earthquakes in southern California the P_d magnitudes agree with the catalog magnitudes with a standard deviation of 0.18 for events less than magnitude 6.5. They concluded that P_d is a robust measurement for estimating the magnitudes of earthquakes for regional early warning purposes in southern California. This method has also applied to Italian region by Zollo et al. [132] with a very good performance.

Because the on-site approach is faster than the regional approach, it can provide useful early warning to sites at short distances from the earthquake epicenter where early warning is most needed. Onsite early warning can be generated by either a single station or by a dense array. For a single station operation, signals from P-waves are used for magnitude and hypocenter determination to predict strong ground shaking. Nakamura [83] first proposed this concept, developed the Urgent Earthquake Detection and Alarm System or UrEDAS [86], and introduced a simple strong-motion index for onsite EEW [85]. However, the reliability of on-site earthquake information is generally less than that obtained with the regional warning system. There currently is a trade-off between warning time and the reliability of the earthquake information. Generally, an information updating procedure is necessary for any EEW system. On-site warning methods can be especially useful in regions where a dense seismic network is deployed.

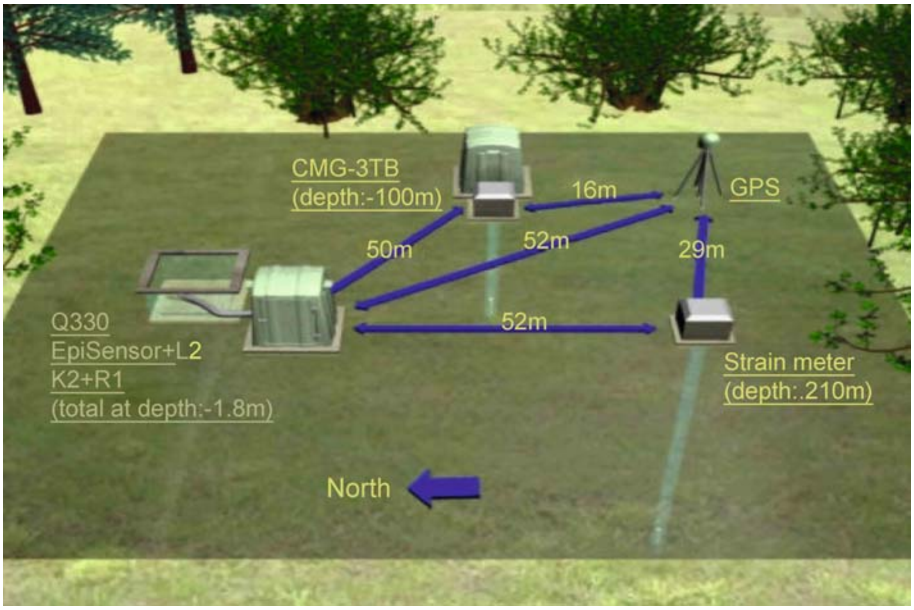
The Japan Meteorological Agency (JMA) began distribution of earthquake early warning information to the public in October 1, 2007 through several means, such as TV and radio [50] (http://www.jma.go.jp/jma/en/Activities/eew.html). The JMA system was successfully activated during the recent Noto Hanto and Niigata Chuetsu–Oki earthquakes in 2007, and provided accurate information of hypocenter, magnitude, and intensity about 3.8 s after the arrival of P-waves at nearby stations. The warning message reached sites further than about 30 km from the epicenter as an early warning alert (i. e., information arrived before shaking started at the site). This is a remarkable performance of the system for damaging earthquakes and gives promise of an early warning system as a practical means for earthquake damage mitigation. Although warning alert is most needed within 30 km of the epicenter, it is not feasible with the current density and configuration of the JMA network.

Lawrence and Cochran [68] proposed a collaborative project for rapid earthquake response and early warning by using the accelerometers that are already installed inside many laptop computers. Their Quake-Catcher Network (QCN) will employ existing laptops, which have accelerometers already installed, and desktops outfitted with inexpensive (under $ 50) USB accelerometers to form the world's largest high-density, distributed computing seismic network for monitoring strong ground motions (http://qcn.stanford.edu/). By freely distributing the necessary software, anyone having a computer with an Internet connection can join the project as a collaborative member. The Quake-Catcher Network also has the potential to provide better understanding of earthquakes, and the client-based software is also intended to be educational, with instructive material displaying the current seismic signal and/or recent earthquakes in the region. It is an effective way to bring earthquake awareness to students and the general public.

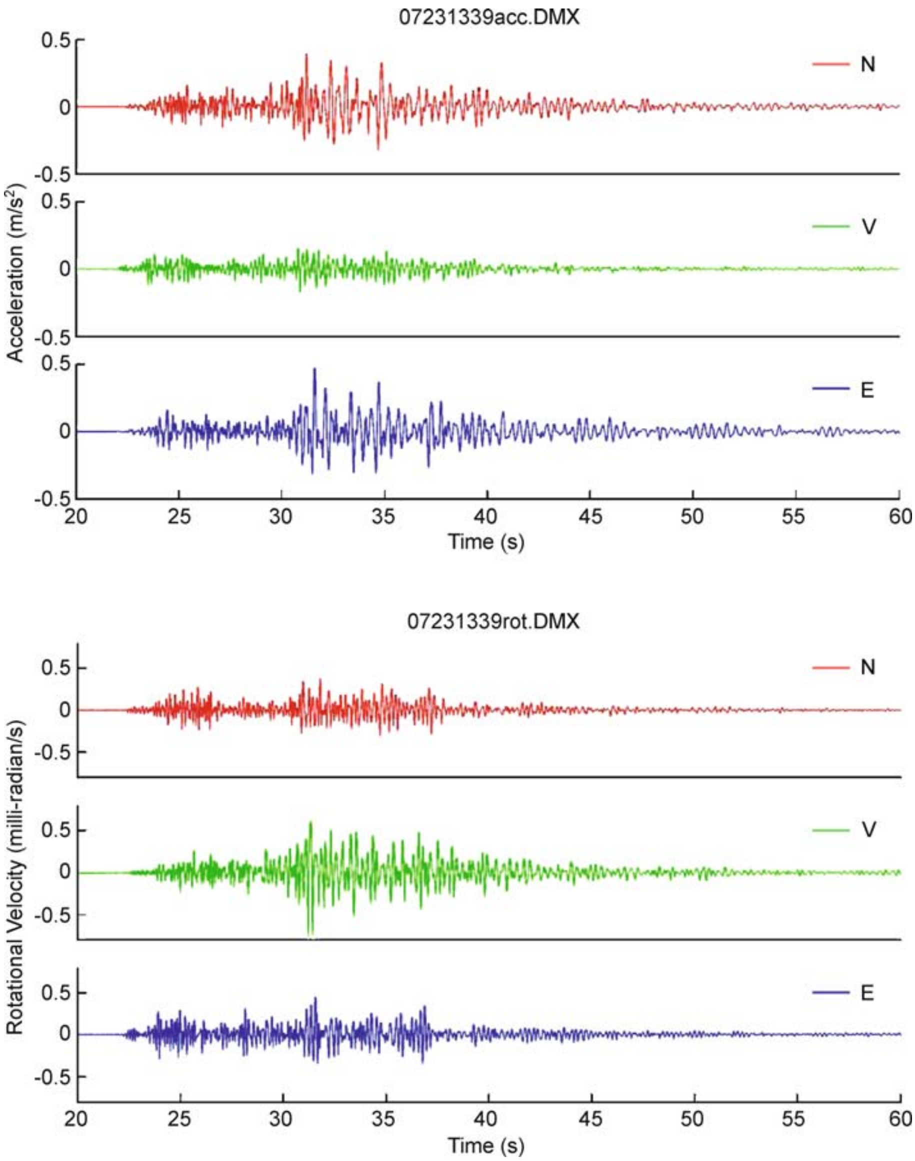
Future Directions

To be successful, monitoring earthquakes requires large, stable funding over a long period of time. The most direct argument for governments to support long-term earthquake monitoring is to collect scientific data for hazard mitigation. In the past two decades about half a million of human lives have been lost due to earthquakes, and economic losses from earthquake damage total about $ 200 billion. Future losses will be even greater as rapid urbanization is taking place worldwide. For example, the recent Japanese Fundamental Seismic Survey and Observation Plan (costing several hundred million US dollars) is a direct response to the economic losses of about $ 100 billion due to the 1995 Kobe earthquake. In addition to scientific and technological challenges in monitoring earthquakes, seismologists must pay attention to achieve (1) stable long-term funding, (2) effective management and execution, and (3) delivery of useful products to the users.

Seismologists benefit greatly from scientific and technological advances in other fields. For example, Global Positioning Systems (GPS) open a new window for monitoring crustal deformation which is important to understand the driving forces that generate earthquakes (► GPS: Applications in Crustal Deformation Monitoring, ► Crustal Deformation During the Seismic Cycle, Interpreting Geodetic Observations of). Under the US Earth Scope Program (http://www.earthscope.org/) the Plate Boundary Observatory (PBO) is covering the western Northern America and Alaska with a network of high precision GPS and strain-meter stations in order to measure deformation across the active boundary between the Pacific and North America plates (http://www.earthscope.org/observatories/pbo). As the sampling rate of GPS data increases, they can provide time histories of displacement during an earthquake. Monitoring earthquakes with multiple types of instruments and sensors is now increasingly popular, and "integrated" or "super" stations are increasingly common. Figure 17 shows an example of an integrated station (HGSD) in eastern Taiwan. Instruments deployed at the HGSD station in eastern Taiwan include: a broadband seismometer, a continuous GPS instrument, a strain-meter, and a 6-channel accelerograph (Model K2 by Kinemetrics) with an internal accelerometer and a rotational sensor (Model R-1 by eentec). A digital seismogram recorded at the HGSD station from an earthquake ($M_W = 5.1$) of July 23, 2007 at a distance of 34 km is



Earthquake Monitoring and Early Warning Systems, Figure 17
Instruments deployed at the HGSD station in eastern Taiwan. Clockwise from the *top*: (1) A broadband seismometer (Model CMG-3TB) installed at a depth of 100 m), (2) A continuous GPS instrument, (3) A strain-meter installed at a depth of 210 m), (4) A Model Q330 6-channel recorder with an accelerometer (Model EpiSensor) and a short-period seismometer (Model L2), and (5) A Model K2 6-channel accelerograph with an internal accelerometer and a rotational sensor (Model R-1)



Earthquake Monitoring and Early Warning Systems, Figure 18
A digital seismogram recorded at the HGSD station from an earthquake ($M_W = 5.1$) of July 23, 2007 at a distance of 34 km. *Top frame*: 3-component translational accelerations. *Bottom frame*: 3-component rotation velocity motions. N = North-South; V = Vertical, and E = East-West

shown in Fig. 18. The importance of rotational seismology and its current status are given in the Appendix.

A radically different design of seismographic networks (and earthquake early warning system in particular) is now possible using the "Sensor Network" developed by Intel Research. Intel is working with the academic community and industry collaborators to actively explore the potential of wireless sensor networks. This research is already demonstrating the potential for this new technology to enhance public safety, reduce the cost of doing business, and bring a host of other benefits to business and society (http://www.intel.com/research/exploratory/wireless_sensors.htm).

It has been very difficult historically to obtain adequate and stable funding for long-term earthquake monitoring, largely because disastrous earthquakes occur infrequently. Since there are many pressing problems facing modern societies, almost all governments react to earthquake (and tsunami) disasters only after the fact, and even then for relatively short periods of time. To advance earthquake prediction research and to develop effective earthquake warning systems will require continuous earthquake monitoring with extensive instrumentations in the near-field for decades and even centuries. Therefore, innovative approaches must be developed and perseverance is needed.

Acknowledgments

We thank John Evans, Fred Klein, Woody Savage, and Chris Stephens for reviewing the manuscript, their comments and suggestions greatly improved it. We are grateful to Lind Gee and Bob Hutt for information about the Global Seismographic Network (GSN) and for providing a high-resolution graphic file of an up-to-date GSN station map.

Appendix: A Progress Report on Rotational Seismology

Seismology is based primarily on the observation and modeling of three orthogonal components of translational ground motions. Although effects of rotational motions due to earthquakes have long been observed (e. g., [80]), Richter (see, p. 213 in [97]) stated that:

> *Perfectly general motion would also involve rotations about three perpendicular axes, and three more instruments for these. Theory indicates, and observation confirms, that such rotations are negligible.*

However, Richter provided no references for this claim, and the available instruments at that time did not have the sensitivity to measure the very small rotation motions that the classical elasticity theory predicts.

Some theoretical seismologists (e. g., [4,5]) and earthquake engineers have argued for decades that the rotational part of ground motions should also be recorded. It is well known that standard seismometers and accelerometers are profoundly sensitive to rotations, particularly tilt, and therefore subject to rotation-induced errors (see e. g., [39,40,41,93]). The paucity of instrumental observations of rotational ground motions is mainly the result of the fact that, until recently, the rotational sensors did not have sufficient resolution to measure small rotational motions due to earthquakes.

Measurement of rotational motions has implications for: (1) recovering the complete ground-displacement history from seismometer recordings; (2) further constraining earthquake rupture properties; (3) extracting information about subsurface properties; and (4) providing additional ground motion information to engineers for seismic design.

In this Appendix, we will first briefly review elastic wave propagation that is based on the linear elasticity theory of simple homogeneous materials under infinitesimal strain. This theory was developed mostly in the early nineteenth century: the differential equations of the linear elastic theory were first derived by Louis Navier in 1821, and Augustin Cauchy gave his formulation in 1822 that remains virtually unchanged to the present day [103]. From this theory, Simeon Poisson demonstrated in 1828 the existence of longitudinal and transverse elastic waves, and in 1885, Lord Rayleigh confirmed the existence of elastic surface waves. George Green put this theory on a physical basis by introducing the concept of strain energy, and, in 1837, derived the basic equations of elasticity from the principle of energy conservation. In 1897, Richard Oldham first identified these three types of waves in seismograms, and linear elasticity theory has been embedded in seismology ever since.

In the following we summarize recent progress in rotational seismology and the need to include measurements of rotational ground motions in earthquake monitoring. The monograph by Teisseyre et al. [109] provides a useful summary of rotational seismology.

Elastic Wave Propagation

The equations of motion for a homogeneous, isotropic, and initially unstressed elastic body may be obtained using the conservation principles of continuum mechanics (e. g., [30]) as

$$\rho \frac{\partial^2 u_i}{\partial t^2} = (\lambda + \mu) \frac{\partial \theta}{\partial x_i} + \mu \nabla^2 u_i , \quad i = 1, 2, 3 \tag{A1}$$

and

$$\theta = \sum_j \partial u_j / \partial x_j \tag{A2}$$

where θ is the dilatation, ρ is the density, u_i is the ith component of the displacement vector $\vec{u}$, t is the time, and λ and μ are the elastic constants of the media. Eq. (A1) may be rewritten in vector form as

$$\rho(\partial^2 \vec{u} / \partial t^2) = (\lambda + \mu) \nabla(\nabla \bullet \vec{u}) + \mu \nabla^2 \vec{u} \, . \tag{A3}$$

If we differentiate both sides of Eq. (A1) with respect to x_i, sum over the three components, and bring ρ to the right-hand side, we obtain

$$\partial^2 \theta / \partial t^2 = [(\lambda + 2\mu)/\rho] \nabla^2 \theta \, . \tag{A4}$$

If we apply the curl operator ($\nabla\times$) to both sides of Eq. (A3), and note that

$$\nabla \bullet (\nabla \times \vec{u}) = 0 \tag{A5}$$

we obtain

$$\partial^2(\nabla \times \vec{u})/\partial t^2 = (\mu/\rho)\nabla^2(\nabla \times \vec{u}) \,. \tag{A6}$$

Now Eqs. (A4) and (A6) are in the form of the classical wave equation

$$\partial^2\Psi/\partial t^2 = v^2\nabla^2\Psi \,, \tag{A7}$$

where Ψ is the wave potential, and v is the wave-propagation velocity (a pseudovector; wave slowness is a proper vector). Thus a dilatational disturbance θ (or a compressional wave) may be transmitted through a homogenous elastic body with a velocity V_P where

$$V_P = \sqrt{[(\lambda + 2\mu)/\rho]} \tag{A8}$$

according to Eq. (A4), and a rotational disturbance $\nabla \times \vec{u}$ (or a shear wave) may be transmitted with a wave velocity V_S where

$$V_S = \sqrt{\mu/\rho} \tag{A9}$$

according to Eq. (A6). In seismology, and for historical reasons, these two types of waves are called the primary (P) and the secondary (S) waves, respectively.

For a heterogeneous, isotropic, and elastic medium, the equation of motion is more complex than Eq. (A3), and is given by Karal and Keller [65] as

$$\begin{aligned}\rho(\partial^2\vec{u}/\partial t^2) &= (\lambda + \mu)\nabla(\nabla \bullet \vec{u}) + \mu\nabla^2\vec{u} \\ &\quad + \nabla\lambda(\nabla \bullet \vec{u}) + \nabla\mu \times (\nabla \times \vec{u}) + 2(\nabla\mu \bullet \nabla)\vec{u} \,.\end{aligned} \tag{A10}$$

Furthermore, the compressional wave motion is no longer purely longitudinal, and the shear wave motion is no longer purely transverse. A review of seismic wave propagation and imaging in complex media may be found in the entry by Igel et al. ▶ Seismic Wave Propagation in Media with Complex Geometries, Simulation of.

A significant portion of seismological research is based on the solution of the elastic wave equations with the appropriate initial and boundary conditions. However, explicit and unique solutions are rare, except for a few simple problems. One approach is to transform the wave equation to the eikonal equation and seek solutions in terms of wave fronts and rays that are valid at high frequencies. Another approach is to develop through specific boundary conditions a solution in terms of normal modes [77]. Although ray theory is only an approximation [17], the classic work of Jeffreys and Bullen, and Gutenberg used it to determine Earth structure and locate earthquakes that occurred in the first half of the 20th century. It remains a principal tool used by seismologists even today. Impressive developments in normal mode and surface wave studies (in both theory and observation) started in the second half of the 20th century, leading to realistic quantification of earthquakes using moment tensor methodology [21].

Rotational Ground Motions

Rotations in ground motion and in structural responses have been deduced indirectly from accelerometer arrays, but such estimates are valid only for long wavelengths compared to the distances between sensors (e. g., [16,34,52,88,90,104]). The rotational components of ground motion have also been estimated theoretically using kinematic source models and linear elastodynamic theory of wave propagation in elastic solids [14,69,70,111].

In the past decade, rotational motions from teleseismic and small local earthquakes were also successfully recorded by sensitive rotational sensors, in Japan, Poland, Germany, New Zealand, and Taiwan (e. g., [53,55,56,105,106,107,108]). The observations in Japan and Taiwan show that the amplitudes of rotations can be *one to two orders of magnitude greater than expected* from the classical linear theory. Theoretical work has also suggested that, in granular materials or cracked continua, asymmetries of the stress and strain fields can create rotations in addition to those predicted by the classical elastodynamic theory for a perfect continuum (▶ Earthquake Source: Asymmetry and Rotation Effects).

Because of lack of instrumentation, rotational motions have not yet been recorded in the near-field (within $\sim$ 25 km of fault ruptures) of strong earthquakes (magnitude > 6.5), where the discrepancy between observations and theoretical predictions may be the largest. Recording such ground motions will require extensive seismic instrumentation along some well-chosen active faults and luck. To this end, several seismologists have been advocating such measurements, and a current deployment in southwestern Taiwan by its Central Weather Bureau is designed to "capture" a repeat of the 1906 Meishan earthquake (magnitude 7.1) with both translational and rotational instruments.

Rotations in structural response, and the contributions to the response from the rotational components of the ground motion, have also been of interest for many decades (e. g., [78,87,98]. Recent reviews on rotational motions in seismology and on the effects of the rotational components of ground motion on structures can be found, for examples, in Cochard et al. [18] and Pillet and Virieux [93], and Trifunac [112], respectively.

Growing Interest – The IWGoRS

Various factors have led to spontaneous organization within the scientific and engineering communities interested in rotational motions. Such factors include: the growing number of successful direct measurements of rotational ground motions (e. g., by ring laser gyros, fiber optic gyros, and sensors based on electro-chemical technology); increasing awareness about the usefulness of the information they provide (e. g., in constraining the earthquake rupture properties, extracting information about subsurface properties, and about deformation of structures during seismic and other excitation); and a greater appreciation for the limitations on information that can be extracted from the translational sensors due to their sensitivity to rotational motions e. g., computation of permanent displacements from accelerograms (e. g., [13,39,40, 41,93,113]).

A small workshop on Rotational Seismology was organized by W.H.K. Lee, K. Hudnut, and J.R. Evans of the USGS on 16 February 2006 in response to grassroots interest. It was held at the USGS offices in Menlo Park and in Pasadena, California, with about 30 participants from about a dozen institutions participating via teleconferencing and telephone [27]. This event led to the formation of the *International Working Group on Rotational Seismology* in 2006, inaugurated at a luncheon during the AGU 2006 Fall Meeting in San Francisco.

The *International Working Group on Rotational Seismology* (IWGoRS) aims to promote investigations of rotational motions and their implications, and the sharing of experience, data, software and results in an open web-based environment (http://www.rotational-seismology.org). It consists of volunteers and has no official status. H. Igel and W.H.K. Lee currently serve as "co-organizers". Its charter is accessible on the IWGoRS web site. The Working Group has a number of active members leading task groups that focus on the organization of workshops and scientific projects, including: testing and verifying rotational sensors, broadband observations with ring laser systems, and developing a field laboratory for rotational motions. The IWGoRS web site also contains the presentations and posters from related meetings, and eventually will provide access to rotational data from many sources.

The IWGoRS organized a special session on *Rotational Motions in Seismology*, convened by H. Igel, W.H.K. Lee, and M. Todorovska during the 2006 AGU Fall Meeting [76]. The goal of that session was to discuss rotational sensors, observations, modeling, theoretical aspects, and potential applications of rotational ground motions. A total of 21 papers were submitted for this session, and over 100 individuals attended the oral session.

The large attendance at this session reflected common interests in rotational motions from a wide range of geophysical disciplines, including strong-motion seismology, exploration geophysics, broadband seismology, earthquake engineering, earthquake physics, seismic instrumentation, seismic hazards, geodesy, and astrophysics, thus confirming the timeliness of IWGoRS. It became apparent that to establish an effective international collaboration within the IWGoRS, a larger workshop was needed to allow sufficient time to discuss the many issues of interest, and to draft research plans for rotational seismology and engineering applications.

First International Workshop

The *First International Workshop on Rotational Seismology and Engineering Applications* was held in Menlo Park, California, on 18–19 September 2007. This workshop was hosted by the US Geological Survey (USGS), which recognized this topic as a new research frontier for enabling a better understanding of the earthquake process and for the reduction of seismic hazards. The technical program consisted of three presentation sessions: plenary (4 papers) and oral (6 papers) held during the first day, and poster (30 papers) held during the morning of the second day. A post-workshop session was held on the morning of September 20, in which scientists of the Laser Interferometer Gravitational-wave Observatory (LIGO) presented their work on seismic isolation of their ultra-high precision facility, which requires very accurate recording of translational and rotational components of ground motions (3 papers). Proceedings of this Workshop were released in Lee et al. [75] with a DVD disc that contains all the presentation files and supplementary information.

One afternoon of the workshop was devoted to in-depth discussions on the key outstanding issues and future directions. The participants could join one of five panels on the following topics: (1) theoretical studies of rotational motions (chaired by L. Knopoff), (2) measuring far-field rotational motions (chaired by H. Igel), (3) measuring near-field rotational motions (chaired by T.L. Teng), (4) engineering applications of rotational motions (chaired by M.D. Trifunac), and (5) instrument design and testing (chaired by J.R. Evans). The panel reports on key issues and unsolved problems, and on research strategies and plans, can be found in Appendices

2.1 through 2.5 in Lee et al. [75]. Following the in-depth group discussions, the panel chairs reported on the group discussions in a common session, with further discussions among all the participants.

Discussions

Since rotational ground motions may play a significant role in the near-field of earthquakes, rotational seismology has emerged as a new frontier of research. During the Workshop discussions, L. Knopoff asked: Is there a quadratic rotation-energy relation, in the spirit of Green's strain-energy relation, coupled to it or independent of it? Can we write a rotation-torque formula analogous to Hooke's law for linear elasticity in the form

$$L_{ij} = d_{ijkl}\omega_{kl} \tag{A11}$$

where ω_{kl} is the rotation,

$$\omega_{kl} = \tfrac{1}{2}(u_{k,l} - u_{l,k})\,. \tag{A12}$$

L_{ij} is the torque density; and d_{ijkl} are the coefficients of rotational elasticity? How are the d's related to the usual c's of elasticity? If we define the rotation vector as

$$\vec{\Omega} = \tfrac{1}{2}(\nabla \times \vec{u}) \tag{A13}$$

we obtain

$$-V_s^2 \nabla \times (\nabla \times \vec{\Omega}) = \partial^2 \vec{\Omega}/\partial t^2 - \tfrac{1}{2}\rho^{-1}(\nabla \times \vec{f}) \tag{A14}$$

where the torque density is $\nabla \times \vec{f}$, $\vec{f}$ is the body force density, and ρ is density of the medium. This shows that rotational waves propagate with S-wave velocity and that it may be possible to store torques. Eq. (15) is essentially an extension using the classical elasticity theory.

Lakes [67] pointed out that the behavior of solids can be represented by a variety of continuum theories. In particular, the elasticity theory of the Cosserat brothers [19] incorporates (1) a local rotation of points as well as the translation motion assumed in the classical theory, and (2) a couple stress (a torque per unit area) as well as the force stress (force per unit area). In the constitutive equation for the classical elasticity theory, there are two independent elastic constants, whereas for the Cosserat elastic theory there are six. Lakes (personal communication, 2007) advocates that there is substantial potential for using generalized continuum theories in geo-mechanics, and any theory must have a strong link with experiment (to determine the constants in the constitutive equation) and with physical reality.

Indeed some steps towards better understandings of rotational motions have taken place. For example, Twiss et al. [114] argued that brittle deformation of the Earth's crust (► Brittle Tectonics: A Non-linear Dynamical System) involving block rotations is comparable to the deformation of a granular material, with fault blocks acting like the grains. They realized the inadequacy of classical continuum mechanics and applied the Cosserat or micropolar continuum theory to take into account two separate scales of motions: macro-motion (large-scale average motion composed of macrostrain rate and macrospin), and micro-motion (local motion composed of microspin). A theoretical link is then established between the kinematics of crustal deformation involving block rotations and the effects on the seismic moment tensor and focal mechanism solutions.

Recognizing that rotational seismology is an emerging field, the *Bulletin of Seismological Society of America* will be publishing in 2009 a special issue under the guest editorship of W.H.K. Lee, M. Çelebi, M.I. Todorovska, and H. Igel.

Bibliography

Primary Literature

1. Ahern TK (2003) The FDSN and IRIS Data Management System: providing easy access to terabytes of information. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part B. Academic Press, Amsterdam pp 1645–1655
2. Aki K (1966) Generation and propagation of G waves from the Niigata Earthquake of June 16, 1964: Part 1. A statistical analysis. Bull Earthq Res Inst 44:23–72; Part 2. Estimation of earthquake moment, released energy, and stress-strain drop from the G wave spectrum. Bull Earthq Res Inst 44:73–88
3. Aki K (1969) Analysis of the seismic coda of local earthquakes as scattered waves. J Geophys Res 74:6215–6231
4. Aki K, Richards PG (1980) Quantitative Seismology. W.H. Freeman, San Francisco
5. Aki K, Richards PG (2002) Quantitative Seismology: Theory and Methods, 2nd edn. University Science Books, Sausalito
6. Allen RM (2007) Earthquake hazard mitigation: New directions and opportunities. In: Kanamori H (ed) Earthquake Seismology. Treatise on Geophysics, vol 4. Elsevier, Amsterdam, pp 607–648
7. Allen RM, Kanamori H (2003) The potential for earthquake early warning in Southern California. Science 300:786–789
8. Ambraseys NN, Jackson JA, Melville CP (2002) Historical seismicity and tectonics: The case of the Eastern Mediterranean and the Middle East. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam, pp 747–763
9. Anderson JG (2003) Strong-motion seismology. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International

Handbook of Earthquake and Engineering Seismology, Part B. Academic Press, Amsterdam pp 937–965
10. Berger J, Davis P, Ekström G (2004) Ambient earth noise: A survey of the Global Seismographic Network. J Geophys Res 109:B11307
11. Borcherdt RD (ed) (1997) Vision for the future of the US National Strong-Motion Program, The committee for the future of the US National Strong Motion Program. US Geol Surv Open-File Rept B97530
12. Bormann P (ed) (2002) New Manual of Seismological Observatory Practice. GeoForschungsZentrum Potsdam http://www.gfz-potsdam.de/bib/nmsop_formular.html
13. Boroschek R, Legrand D (2006) Tilt motion effects on the double-time integration of linear accelerometers: an experimental approach. Bull Seism Soc Am 96:2072–2089
14. Bouchon M, Aki K (1982) Strain, tilt, and rotation associated with strong ground motion in the vicinity of earthquake faults. Bull Seism Soc Am 72:1717–1738
15. Burridge R, Knopoff L (1964) Body force equivalents for seismic dislocations. Bull Seism Soc Am 54:1875–1888
16. Castellani A, Boffi G (1986) Rotational components of the surface ground motion during an earthquake. Earthq Eng Struct Dyn 14:751–767
17. Chapman CH (2002) Seismic ray theory and finite frequency extensions. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam, pp 103–123
18. Cochard A, Igel H, Schuberth B, Suryanto W, Velikoseltsev A, Schreiber U, Wassermann J, Scherbaum F, Vollmer D (2006) Rotational motions in seismology: theory, observation, simulation. In: Teisseyre R, Takeo M, Majewski M (eds) Earthquake Source Asymmetry, Structural Media and Rotation Effects. Springer, Heidelberg, pp 391–411
19. Cosserat E, Cosserat F (1909) Theorie des Corps Deformables. Hermann, Paris
20. Dziewonski AM, Woodhouse JH (1983) An experiment in the systematic study of global seismicity: centroid-moment tensor solutions for 201 moderate and large earthquakes of 1981. J Geophys Res 88:3247–3271
21. Dziewonski AM, Woodhouse JH (1983) Studies of the seismic source using normal-mode theory. In: Kanamori H, Boschi B (eds) Earthquakes: Observation, Theory, and Interpretation. North-Holland, Amsterdam, pp 45–137
22. Dziewonski AM, Chou TA, Woodhouse JH (1981) Determination of earthquake source parameters from waveform data for studies of global and regional seismicity. J Geophys Res 86:2825–2852
23. Ekström G (1994) Rapid earthquake analysis utilizes the Internet. Comput Phys 8:632–638
24. Engdahl ER, Villasenor A (2002) Global seismicity: 1900–1999. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam, pp 665–690
25. Espinosa-Aranda JM, Jimenez A, Ibarrola G, Alcantar F, Aguilar A, Inostroza M, Maldonado S (1995) Mexico City Seismic Alert System. Seism Res Lett 66(6):42–53
26. Evans JR, Hamstra RH, Kündig C, Camina P, Rogers JA (2005) TREMOR: A wireless MEMS accelerograph for dense arrays. Earthq Spectr 21(1):91–124
27. Evans JR, Cochard A, Graizer V, Huang B-S, Hudnut KW, Hutt CR, Igel H, Lee WHK, Liu C-C, Majewski E, Nigbor R, Safak E, Savage WU, Schreiber U, Teisseyre R, Trifunac M, Wassermann J, Wu C-F(2007) Report of a workshop on rotational ground motion. US Geol Surv Open File Rep 20:2007–1145 http://pubs.usgs.gov/of/2007/1145/
28. Feigl KL (2002) Estimating earthquake source parameters from geodetic measurements. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam, pp 607–620
29. Field EH, Hough SE (1997) The variability of PSV response spectra across a dense array deployed during the Northridge aftershock sequence. Earthq Spectr 13:243–257
30. Fung YC (1965) Foundations of Solid Mechanics. Prentice-Hall, Englewood Cliffs
31. Gasparini P, Manfredi G, Zschau J (eds) (2007) Seismic Early Warning Systems. Springer, Berlin
32. Gee L, Neuhauser D, Dreger D, Pasyanos M, Uhrhammer R, Romanowicz B (2003) The Rapid Earthquake Data Integration Project. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part B. Academic Press, Amsterdam, pp 1261–1273
33. Geiger LC (1912) Probability method for the determination of earthquake epicenters from the arrival time only. Bull St Louis Univ 8:60–71
34. Ghayamghamian MR, Nouri GR (2007) On the characteristics of ground motion rotational components using Chiba dense array data. Earthq Eng Struct Dyn 36(10):1407–1429
35. Gilbert F (1971) Excitation of the normal modes of the Earth by earthquake sources. Geophys J R Astron Soc 22:223–226
36. Gilbert F, Dziewonski AM (1975) Application of normal mode theory to the retrieval of structural parameters and source mechanisms from seismic spectra. Phil Trans Roy Soc Lond A 278:187–269
37. Goltz JD, Flores PJ, Chang SE, Atsumi T (2001) Emergency response and early recovery. In: 1999 Chi-Chi, Taiwan, Earthquake Reconnaissance Report. Earthq Spectra Suppl A 17:173–183
38. Goodstein JR (1991) Millikan's School: A History of the California Institute of Technology. Norton, New York
39. Graizer VM (1991) Inertial seismometry methods. Izv Earth Phys Akad Nauk SSSR 27(1):51–61
40. Graizer VM (2005) Effect of tilt on strong motion data processing. Soil Dyn Earthq Eng 25:197–204
41. Graizer VM (2006) Tilts in strong ground motion. Bull Seis Soc Am 96:2090–2106
42. Guidoboni E (2002) Historical seismology: the long memory of the inhabited world. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam, pp 775–790
43. Gutenberg B (1945) Amplitudes of surface waves and magnitudes of shallow earthquakes. Bull Seism Soc Am 35:3–12
44. Gutenberg B, Richter CF (1954) Seismicity of the Earth, 2nd edn. Princeton University Press, Princeton
45. Gutenberg B, Richter CF (1956) Magnitude and energy of earthquakes. Ann Geofis 9:1–15

46. Hanks TC, Wyss M (1972) The use of body wave spectra in the determination of seismic source parameters. Bull Seism Soc Am 62:561–589
47. Hauksson E, Jones LM, Shakal AF (2003) TriNet: a modern ground motion seismic network. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part B. Academic Press, Amsterdam, pp 1275–1284
48. Havskov J, Alguacil G (2004) Instrumentation in Earthquake Seismology. Springer, Berlin
49. Horiuchi S, Negishi H, Abe K, Kamimura A, Fujinawa Y (2005) An automatic processing system for broadcasting earthquake alarms. Bull Seism Soc Am 95:708–718
50. Hoshiba M, Kamigaichi O, Saito M, Tsukada S, Hamada N (2008) Earthquake early warning starts nationwide in Japan, EOS. Trans Am Geophys Un 89(8):73–74
51. Housner GW (2002) Historical view of earthquake engineering. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam, pp 13–18
52. Huang BS (2003) Ground rotational motions of the 1991 Chi-Chi, Taiwan earthquake as inferred from dense array observations. Geophys Res Lett 30(6):1307–1310
53. Huang BS, Liu CC, Lin CR, Wu CF, Lee WHK (2006) Measuring mid- and near-field rotational ground motions in Taiwan. Poster, presented at 2006 Fall AGU Meeting. San Francisco
54. Hutt CR, Bolton HF, Holcomb LG (2002) US contribution to digital global seismograph networks. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam, pp 319–322
55. Igel H, Cochard A, Wassermann J, Schreiber U, Velikoseltsev A, Dinh NP (2007) Broadband observations of rotational ground motions. Geophys J Int 168(1):182–197
56. Igel H, Schreiber U, Flaws A, Schuberth B, Velikoseltsev A, Cochard A (2005) Rotational motions induced by the M8.1 Tokachi-oki earthquake, September 25, 2003. Geophys Res Lett 32:L08309. doi:10.1029/2004GL022336
57. Kamigaichi O (2004) JMA Earthquake Early Warning. J Japan Assoc Earthq Eng 4:134–137
58. Kanamori H (1977) The energy release in great earthquakes. J Geophys Res 82:2921–2987
59. Kanamori H (1978) Quantification of earthquakes. Nature 271:411–414
60. Kanamori H (2003) Earthquake prediction: an overview. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part B. Academic Press, Amsterdam, pp 1205–1216
61. Kanamori H (2005) Real-time seismology and earthquake damage mitigation. Annual Rev Earth Planet Sci 33:195–214
62. Kanamori H, Brodsky EE (2000) The physics of earthquakes. Phys Today 54(6):34–40
63. Kanamori H, Rivera L (2006) Energy partitioning during an earthquake. In: Abercrombie R, McGarr A, Kanamori H, Di Toro G (eds) Earthquakes: Radiated Energy and the Physics of Faulting. Geophysical Monograph, vol 170. Am Geophys Union, Washington DC, pp 3–13
64. Kanamori H, Hauksson E, Heaton T (1997) Real-time seismology and earthquake hazard mitigation. Nature 390:461–464
65. Karal FC, Keller JB (1959) Elastic wave propagation in homogeneous and inhomogeneous media. J Acoust Soc Am 31:694–705
66. Kisslinger C, Howell BF (2003) Seismology and physics of the Earth's interior in the USA, 1900–1960. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part B. Academic Press, Amsterdam, p 1453
67. Lakes RS (1995) Experimental methods for study of Cosserat elastic solids and other generalized continua. In: Mühlhaus H (ed) Continuum Models for Materials with Micro-structure. Wiley, New York, pp 1–22
68. Lawrence JF, Cochran ES (2007) The Quake Catcher Network: Cyberinfrastructure bringing seismology into schools and homes. American Geophysical Union, Fall Meeting 2007, abstract #ED11C-0633
69. Lee VW, Trifunac MD (1985) Torsional accelerograms. Int J Soil Dyn Earthq Eng 4(3):132–139
70. Lee VW, Trifunac MD (1987) Rocking strong earthquake accelerations. Int J Soil Dyn Earthq Eng 6(2):75–89
71. Lee WHK (2002) Challenges in observational seismology. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam, 269–281
72. Lee WHK, Benson RB (2008) Making non-digitally-recorded seismograms accessible online for studying earthquakes. In: Fréchet J, Meghraoui M, Stucchi M (eds) Modern Approach in Historical Seismology: Interdisciplinary studies of past and recent earthquakes. Springer, Berlin, pp 403–427
73. Lee WHK, Espinosa-Aranda JM (2003) Earthquake early warning systems: Current status and perspectives. In: Zschau J, Kuppers AN (eds) Early Warning Systems for Natural Disaster, Reduction. Springer, Berlin, pp 409–423
74. Lee WHK, Stewart SW (1981) Principles and Applications of Microearthquake Networks. Academic Press, New York
75. Lee WHK, Celebi M, Todorovska MI, Diggles MF (2007) Rotational seismology and engineering applications: Proceedings for the First International Workshop, Menlo Park, California, USA, 18–19 September. US Geol Surv Open File Rep 2007–1144. http://pubs.usgs.gov/of/2007/1144/
76. Lee WHK, Igel H, Todorovska MI, Evans JR (2007) Rotational Seismology: AGU Session, Working Group, and Website. US Geol Surv Open File Rep 2007–1263. http://pubs.usgs.gov/of/2007/1263/
77. Lognonne P, Clevede E (2002) Normal modes of the Earth and planets. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam, pp 125–147
78. Luco JE (1976) Torsional response of structures to obliquely incident seismic SH waves. Earthq Eng Struct Dyn 4:207–219
79. Mallet R (1858) Fourth report on the facts of earthquake phenomena. Ann Rep Brit Assn Adv Sci 28:1–136
80. Mallet R (1862) Great Neapolitan Earthquake of 1857, vol I, II. Chapman and Hall, London
81. Maruyama T (1963) On the force equivalents of dynamical elastic dislocations with reference to the earthquake mechanism. Bull Earthq Res Inst 41:467–486
82. Michelini A, De Simoni B, Amato A, Boschi E (2005) Collecting, digitizing and distributing historical seismological data. EOS 86(28) 12 July 2005

83. Nakamura Y (1984) Development of the earthquake early-warning system for the Shinkansen, some recent earthquake engineering research and practical in Japan. The Japanese National Committee of the International Association for Earthquake Engineering, pp 224–238
84. Nakamura Y (1988) On the urgent earthquake detection and alarm system, UrEDAS. Proc Ninth World Conf Earthq Eng 7:673–678
85. Nakamura Y (2004) On a rational strong motion index compared with other various indices. 13th World Conf Earthq Eng, Paper No 910
86. Nakamura Y, Saita J (2007) UrEDAS, The Earthquake Warning System: today and tomorrow. In: Gasparini P, Manfredi G, Zschau J (eds) Earthquake Early Warning Systems. Springer, Berlin, pp 249–281
87. Newmark NM (1969) Torsion in symmetrical buildings. Proc. Fourth World Conference on Earthquake Eng, vol II. pp A3/19–A3/32
88. Niazi M (1987) Inferred displacements, velocities and rotations of a long rigid foundation located at El-Centro differential array site during the 1979 Imperial Valley, California, earthquake. Earthq Eng Struct Dyn 14:531–542
89. Odaka T, Ashiya K, Tsukada S, Sato S, Ohtake K, Nozaka D (2003) A new method of quickly estimating epicentral distance and magnitude from a single seismic record. Bull Seism Soc Am 93:526–532
90. Oliveira CS, Bolt BA (1989) Rotational components of surface strong ground motion. Earthq Eng Struct Dyn 18:517–526
91. Oliver J, Murphy L (1971) WWNSS: Seismology's global network of observing stations. Science 174:254–261
92. Peterson J (1993) Observations and modeling of seismic background noise. US Geol Surv Open File Rep, 93–322
93. Pillet R, Virieux J (2007) The effects of seismic rotations on inertial sensors. Geophys J Int. doi:10.1111/j.1365-246X.2007.03617.x
94. Reid HF (1910) The California Earthquake of 18 April 1906, vol 2. The Mechanics of the Earthquake. Carnegie Inst, Washington DC
95. Richards PG (2002) Seismological methods of monitoring compliance with the Comprehensive Nuclear Test-Ban Treaty. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam, pp 369–382
96. Richter CF (1935) An instrumental earthquake magnitude scale. Bull Seis Soc Am 25:1–32
97. Richter CF (1958) Elementary Seismology. Freeman, San Francisco
98. Rutenberg A, Heidebrecht AC (1985) Rotational ground motion and seismic codes. Can J Civ Eng 12(3):583–592
99. Saita J, Nakamura Y (2003) UrEDAS: the early warning system for mitigation of disasters caused by earthquakes and tsunamis. In: Zschau J, Kuppers AN (eds) Early Warning Systems for Natural Disaster, Reduction. Springer, Berlin, pp 453–460
100. Savage JC (1978) Dislocation in seismology. In: Nabarro FRN (ed) Dislocation in Solids. North-Holland, Amsterdam, 251–339
101. Schweitzer J, Lee WHK (2003) Old seismic bulletins: a collective heritage from early seismologists. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part B. Academic Press, Amsterdam, pp 1665–1717 (with CD-ROM)
102. Shin TC, Tsai YB, Yeh YT, Liu CC, Wu YM (2003) Strong-motion instrumentation programs in Taiwan. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part B. Academic Press, Amsterdam, pp 1057–1062
103. Sokolnikoff IS (1956) Mathematical Theory of Elasticity, 2nd edn. McGraw-Hill, New York
104. Spudich P, Steck LK, Hellweg M, Fletcher JB, Baker LM (1995) Transient stresses at Park-field, California, produced by the m 7.4 Landers earthquake of 28 June 1992: Observations from the UPSAR dense seismograph array. J Geophys Res 100:675–690
105. Suryanto W, Igel H, Wassermann J, Cochard A, Schubert B, Vollmer D, Scherbaum F (2006) Comparison of seismic array-derived rotational motions with direct ring laser measurements. Bull Seism Soc Am 96(6):2059–2071
106. Takeo M (1998) Ground rotational motions recorded in near-source region. Geophy Res Lett 25(6):789–792
107. Takeo M, Ito HM (1997) What can be learned from rotational motions excited by earthquakes? Geophys J Int 129:319–329
108. Teisseyre R, Suchcicki J, Teisseyre KP, Wiszniowski J, Palangio P (2003) Seismic rotation waves: basic elements of theory and recording. Annali Geofis 46:671–685
109. Teisseyre R, Takeo M, Majewski E (eds) (2006) Earthquake Source Asymmetry, Structural Media and Rotation Effects. Springer, Berlin
110. Teng TL, Wu L, Shin TC, Tsai YB, Lee WHK (1997) One minute after: strong motion map, effective epicenter, and effective magnitude. Bull Seism Soc Am 87:1209–1219
111. Trifunac MD (1982) A note on rotational components of earthquake motions on ground surface for incident body waves. Soil Dyn Earthq Eng 1:11–19
112. Trifunac MD (2006) Effects of torsional and rocking excitations on the response of structures. In: Teisseyre R, Takeo M, Majewski M (eds) Earthquake Source Asymmetry, Structural Media and Rotation Effects. Springer, Heidelberg, pp 569–582
113. Trifunac MD, Todorovska MI (2001) A note on the useable dynamic range of accelerographs recording translation. Soil Dyn Earthq Eng 21(4):275–286
114. Twiss R, Souter B, Unruh J (1993) The effect of block rotations on the global seismic moment tensor and the patterns of seismic P and T axes. J Geophys Res 98(B1):645–674
115. Utsu T (2002) A list of deadly earthquakes in the world (1500–2000). In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam, pp 691–717
116. Utsu T (2002) Relationships between magnitude scales. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam, pp 733–746
117. Uyeda S (2002) Continental drift, sea-floor spreading, and plate/plume tectonics. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam, pp 51–67
118. Villasenor A, Engdahl ER (2007) Systematic relocation of early instrumental seismicity: Earthquakes in the International Seismological Summary for 1960–1963. Bull Seism Soc Am 97:1820–1832

119. Wald DJ, Quitoriano V, Heaton TH, Kanamori H (1999) Relationships between peak ground acceleration, peak ground velocity, and modified Mercalli intensity in California. Earthq Spectr 15:557–564
120. Wald DJ, Quitoriano V, Heaton TH, Kanamori H, Scrivner CW, Worden CB (1999) TriNet "ShakeMaps": Rapid generation of peak ground motion and intensity maps for earthquakes in Southern California. Earthq Spectr 15:537–555
121. Webb SC (2002) Seismic noise on land and on the seafloor. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam, pp 305–318
122. Wielandt E (2002) Seismometry. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, Amsterdam, pp 283–304
123. Willemann RJ, Storchak DA (2001) Data collection at the International Seismological Centre. Seism Res Lett 72:440–453
124. Wu YM, Kanamori H (2005) Experiment on an onsite early warning method for the Taiwan early warning system. Bull Seism Soc Am 95:347–353
125. Wu YM, Kanamori H (2005) Rapid assessment of damaging potential of earthquakes in Taiwan from the beginning of P Waves. Bull Seism Soc Am 95:1181–1185
126. Wu YM, Kanamori H (2008) Exploring the feasibility of on-site earthquake early warning using close-in records of the 2007 Noto Hanto earthquake. Earth Planets Space 60:155–160
127. Wu YM, Zhao L (2006) Magnitude estimation using the first three seconds P-wave amplitude in earthquake early warning. Geophys Res Lett 33:L16312. doi:10.1029/2006GL026871
128. Wu YM, Chen CC, Shin TC, Tsai YB, Lee WHK,Teng TL (1997) Taiwan Rapid Earthquake Information Release System. Seism Res Lett 68:931–943
129. Wu YM, Hsiao NC, Lee WHK, Teng TL, Shin TC (2007) State of the art and progresses of early warning system in Taiwan. In: Gasparini P, Manfredi G, Zschau J (eds) Earthquake Early Warning Systems. Springer, Berlin, pp 283–306
130. Wu YM, Kanamori H, Allen R, Hauksson E (2007) Determination of earthquake early warning parameters, τ_c and P_d, for southern California. Geophys J Int 169:667–674
131. Wu YM, Lee WHK, Chen CC, Shin TC, Teng TL, Tsai YB (2000) Performance of the Taiwan Rapid Earthquake Information Release System (RTD) during the 1999 Chi-Chi (Taiwan) earthquake. Seism Res Lett 71:338–343
132. Zollo A, Lancieri M, Nielsen S (2006) Earthquake magnitude estimation from peak amplitudes of very early seismic signals on strong motion records. Geophys Res Lett 33:L23312. doi:10.1029/2006GL027795

Books and Reviews

Abercrombie R, McGarr A, Kanamori H, Di Toro G (2006) Earthquakes: Radiated Energy and the Physics of Faulting. Geophysical Monograph, vol 170. American Geophysical Union, Washington DC
Bolt BA (1993) Earthquakes. W.H. Freeman, New York
Chen YT, Panza GF, Wu ZL (2004) Earthquake Hazard, Risk, and Strong Ground Motion. Seismological Press, Beijing
Kanamori H (ed) (2007) Earthquake Seismology, Treatise on Geophysics, vol 4. Elsevier, Amsterdam
Kanamori H, Boschi E (1983) Earthquakes: Observation, Theory and Interpretation. North-Holland, Amsterdam
Keilis-Borok VI, Soloviev AA (eds) (2003) Nonlinear Dynamics of the Lithosphere and Earthquake Prediction. Springer, Berlin
Lee WHK, Meyers H, Shimazaki K (eds) (1988) Historical Seismograms and Earthquakes of the World. Academic Press, San Diego
Lee WHK, Kanamori H, Jennings JC, Kisslinger C (eds) (2002) International Handbook of Earthquake and Engineering Seismology, Part A. Academic Press, San Diego, pp 933 (and 1 CD-ROM)
Lee WHK, Kanamori H, Jennings JC, Kisslinger C (eds) (2003) International Handbook of Earthquake and Engineering Seismology, Part B. Academic Press, San Diego, pp 1009 (and 2 CD-ROMs)
Pujol J (2003) Elastic Wave Propagation and Generation in Seismology. Cambridge Univ Press, Cambridge
Zschau J, Kuppers AN (eds) (2003) Early Warning Systems for Natural Disaster Reduction. Springer, Berlin

Earthquake Networks, Complex

Sumiyoshi Abe[1,2], Norikazu Suzuki[3]
[1] Department of Physical Engineering, Mie University, Tsu, Japan
[2] Institut Supérieur des Matériaux et Mécaniques, Le Mans, France
[3] College of Science and Technology, Nihon University, Chiba, Japan

Article Outline

Glossary

Network or graph A network (or a graph) [28] consists of vertices (or nodes) and edges (or links) connecting them. In general, a network contains loops (i. e., edges with both ends attached to the same vertices) and multiple edges (i. e., edges more than one that connect two different vertices). If edges have their directions,