# Chapter 1

# THE NATURE OF EARTHQUAKE GROUND MOTION

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Key words: Attenuation, Causes of Earthquakes, Collapse Earthquakes, Damage Mechanisms, Directivity, Earthquakes, Earthquake Faults, Earthquake Prediction, Elastic Rebound, Fault Types, Intensity Scales, Intraplate Earthquakes, Magnitude, Magnitude Scales, Magnitude Saturation, MMI, Near-Fault Effects, Plate Tectonics, Reservoir-Induced Earthquakes, Seismicity, Seismic Moment, Seismic Risk, Seismic Waves, Seismology, Source Models, Strong Ground Motion, Surface Rupture, Tectonic Earthquakes, Volcanic Earthquakes

Abstract: The aim of this chapter is to provide a basic understanding about earthquakes, their world-wide distribution, what causes them, their likely damage mechanisms, earthquake measuring scales, and current efforts on the prediction of strong seismic ground motions. This chapter, therefore, furnishes the basic information necessary for understanding the more detailed concepts that follow in the subsequent chapters of this book. The basic vocabulary of seismology is defined. The seismicity of the world is discussed first and its relationship with tectonic plates is explained. The general causes of earthquakes are discussed next where tectonic actions, dilatancy in the crustal rocks, explosions, collapses, volcanic actions, and other likely causes are introduced. Earthquake fault sources are discussed next. Various faulting mechanisms are explained followed by a brief discussion of seismic waves. Earthquake damage mechanisms are introduced and different major damage mechanisms are identified by examples. Quantification of earthquakes is of significant interest to seismic design engineers. Various earthquake intensity and magnitude scales are defined followed by a description of earthquake source models. Basic information regarding the concepts of directivity and near-fault effects are presented. Finally, the ideas behind seismic risk evaluation and earthquake prediction are discussed.

### **1.1 INTRODUCTION**

On the average, 10,000 people die each year from earthquakes (see Figure 1-1). A UNESCO study gives damage losses amounting to \$10,000,000,000 from 1926 to 1950 from earthquakes. In Central Asia in this interval two towns and 200 villages were destroyed. Since then several towns including Ashkhabad Agadir (1960), Skopje (1963), (1948),Managua (1972), Gemona (1976), Tangshan (1976), Mexico City (1985), Spitak (1988), Kobe (1995), cities in Turkey and Taiwan (1999) and hundreds of villages have been damaged ground severely by shaking. Historical writings testify to man's long concern about earthquake hazards.

The first modern stimulus for scientific study of earthquakes came from the extensive field work of the Irish engineer, Robert Mallett, after the great Neopolitan earthquake of 1857 in southern Italy. He set out to explain the "masses of dislocated stone and mortar" in terms of mechanical principles and in doing so established basic vocabulary such as *seismology*, *hypocenter* and *isoseismal*. Such close links between engineering and seismology have continued ever since<sup>(1-1,1-2)</sup>.

It is part of strong motion seismology to explain and predict the large amplitude-long duration shaking observed in damaging earthquakes. In the first sixty years of the century, however, the great seismological advances occurred in studying waves from distant earthquakes using very sensitive seismographs. Because the wave amplitudes in even a nearby magnitude 5 earthquake would exceed the dynamic range of the usual seismographs, not much fundamental work was done by seismologists on the rarer large earthquakes of engineering importance.

Nowadays, the situation has changed. After the 1971 San Fernando earthquake, hundreds of strong-motion records were available for this magnitude 6.5 earthquake. The 1.2g recorded at Pacoima Dam led to questions on topographic



Figure 1-1. Loss of life caused by major earthquakes [After Hiroo Kanamori<sup>(1-10)</sup>].

amplification and the construction of realistic models of fault-rupture and travel-path that could explain the strong motion patterns. Progress on these seismological questions followed rapidly in studies of variation in ground motions in the 1989 Loma Prieta earthquake (M 7.0), the 1994 Northridge earthquake (M 6.8) and the 1999 Chi Chi event in Taiwan (M 7.6). A harvest of strong motion recordings was obtained in the latter earthquake, showing numerous horizontal peak accelerations in the range 0.5g to 1.0g. Digital recorders and fast computers mean that both seismologists and engineers can tackle more fundamental and realistic problems of earthquake generation and ground shaking.

Knowledge of strong ground shaking is now advancing rapidly, largely because of the growth of appropriately sited strong-motion accelerographs in seismic areas of the world. For example, in the Strong Motion Instrumentation Program in California, by the year 2000 there were 800 instruments in the free-field and 130 buildings and 45 other structures instrumented. Over 500 records had been digitized and were available for use in research or practice (see Chapter 16). In earthquake-prone regions, structural design of large or critical engineered structures such as high-rise buildings, large dams, and bridges now usually involves quantitative dynamic analysis; engineers ask penetrating questions on the likely seismic intensity for construction sites and require input motions or spectra of defining parameters. Predicted seismograms (time-histories) for dynamic modeling in structural design or vulnerability assessments are often needed.

The aim of this chapter is to provide a basic understanding about earthquakes, their worldwide distribution, what causes them, their likely damage mechanisms, earthquake measuring scales, and current efforts on the prediction of strong seismic ground motions. Additional helpful background on the subject may be found in References 1-2 through 1-34.

# 1.2

# SEISMICITY OF THE WORLD

From the earthquake wave readings at different seismographic observatories, the position of the center of an earthquake can be calculated<sup>(1-1)</sup>. In this way, a uniform picture of earthquake distribution around the world has been obtained (see Figure 1-2). Definite belts of seismic activity separate large oceanic and continental regions, themselves mainly, but by no means completely, devoid of earthquake centers. Other concentrations of earthquake sources can be seen in the oceanic areas, for example, along the center of the Atlantic and Indian Oceans. These are the sites of gigantic submarine mountain ranges called mid-oceanic ridges. The geological strains that prevail throughout this global ridge system are evidenced by mountain peaks and deep rift valleys. Volcanic eruptions are frequent, and earthquakes originating along these ridges often occur in swarms, so that many hundreds of shocks are concentrated in a small area in a short time.

Dense concentrations of earthquake centers with some as much as 680 kilometers beneath the surface also coincide with island arcs, such as those of the Pacific and the eastern Caribbean.

On the western side of the Pacific Ocean, the whole coast of Central and South America is agitated by many earthquakes, great and small. High death tolls have ensued from the major ones. In marked contrast, the eastern part of South America is almost entirely free from earthquakes, and can be cited as an example of low seismic risk country. Other seismically quiet continental areas can be seen in Figure 1-2.

In Europe, earthquake activity is quite widespread. To the south, Turkey, Greece, Yugoslavia, Italy, Spain and Portugal suffer from it, and large numbers of people have died in disasters throughout the years. An earthquake off southwest Iberia on November 1, 1755 produced a great tsunami, which caused many of the 50,000 to 70,000 deaths



*Figure 1-2.* Tectonic plates and world-wide distribution of earthquakes. (From *Earthquakes*, by Bruce A. Bolt. Copyright 1978, 1999 W. H. Freeman and Company. Used with permission.)

occurring in Lisbon, Portugal, and surrounding areas; the shaking was felt in Germany and the Netherlands. In Alicante, Spain, on March 21, 1829, a shock killed about 840 persons and injured many hundred more. Total or partial destruction of more than 5,000 houses was reported in and near Torrevieja and Murcia. On December 28, 1908, a devastating earthquake hit Messina, Italy, causing 120,000 deaths and widespread damage. The most recent deadly one to affect that country struck on May 6, 1976, in the Friuli region near Gemona; about 965 persons were killed and 2280 injured.

On December 27, 1939, in Erzincan, Turkey, 23,000 lives were lost from a major earthquake. Similar killer earthquakes have occurred in Turkey and Iran in recent years. The Erzincan earthquake along the Anatolian fault in Turkey on March 13, 1992 caused many building collapses and the June 21, 1990 earthquake (M 7.3) devastated two Iranian provinces, Gilan and Zanjan. August 17, 1999 saw a 50 km rupture of the north Anatoliam fault along the Marmara Sea south of Izmit producing a magnitude 7.4 earthquake and over 16,000 deaths.

North of the Mediterranean margin, Europe is much more stable. However, destructive earthquakes do occur from time to time in Romania, Germany, Austria and Switzerland, and even in the North Sea region and Scandinavia. For example, on October 8, 1927, an earthquake occurred near Schwadorf in Austria and caused damage in an area southeast of Vienna. This earthquake was felt in Hungary, Germany, and Czechoslovakia at distances of 250 kilometers from the center of the disturbance. The seismicity in the North Sea is sufficiently significant to require attention to earthquake resistant design of oil platforms there.

In Africa, damaging earthquakes have occurred in historical times. A notable case was the magnitude 5.6 earthquake on November 14, 1981 that was felt in Aswan, Egypt. This earthquake was probably stimulated by the impounding of water in Lake Nassar behind the high Aswan Dam.

An example of infrequent and dispersed seismicity is the occurrence of earthquakes in Australia. Nevertheless, this country does have some significant areas of present-day seismicity. Of particular interest is a damaging earthquake of moderate size that was centered near Newcastle and causing major damage and killing fourteen people. It was a surprise from a seismological point of view because no fault maps were available which showed seismogenic geological structures near Newcastle.

During an earthquake, seismic waves radiate from the earthquake source somewhere below the ground surface as opposite sides of a slipping fault rebound in opposite directions in order to decrease the strain energy in the rocks. Although in natural earthquakes this source is spread out through a volume of rock, it is often convenient to imagine a simplified earthquake source as a point from which the waves first emanate. This point is called the *earthquake focus*. The point on the ground surface directly above the focus is called the *earthquake epicenter*.

Although many foci are situated at shallow depths, in some regions they are hundreds of kilometers deep. Such regions include the South American Andes, the Tonga Islands, Samoa, the New Hebrides chain, the Japan Sea, Indonesia, and the Caribbean Antilles. On the average, the frequency of occurrence of earthquakes in these regions declines rapidly below a depth of 200 kilometers, but some foci are as deep as 680 kilometers. Rather arbitrarily, earthquakes with foci from 70 to 300 kilometers deep are called intermediate focus and those below this depth are termed deep focus. Some intermediate and deep focus earthquakes are located away from the Pacific region, in the Hindu Kush, in Romania, in the Aegean Sea and under Spain.

The shallow-focus earthquakes (focus depth less than 70 kilometers) wreak the most devastation, and they contribute about three quarters of the total energy released in earthquakes throughout the world. In California, for example, all of the known earthquakes to date have been shallow-focus. In fact, it has been shown that the great majority of earthquakes occurring in central California originate from foci in the upper five kilometers of the Earth, and only a few are as deep as even 15 kilometers.

Most moderate to large shallow earthquakes are followed, in the ensuing hours and even in the next several months, by numerous, usually smaller earthquakes in the same vicinity. These earthquakes are called *aftershocks*, and large earthquakes are sometimes followed by incredible numbers of them. The great Rat Island earthquake in the Aleutian Island on February 4, 1965 was, within the next 24 days, followed by more than 750 aftershocks large enough to be recorded by distant seismographs. Aftershocks are sometimes energetic enough to cause additional damage to already weakened structures. This happened, for example, a week after the Northridge earthquake of January 17, 1994 in the San Fernando Valley when some weakened structures sustained additional cracking from magnitude 5.5 aftershocks. A few earthquakes are preceded by smaller foreshocks from the source area, and it has been suggested that these can be used to predict the main shock.

# 1.3 CAUSES OF EARTHQUAKES

#### **1.3.1** Tectonic Earthquakes

In the time of the Greeks it was natural to link the Aegean volcanoes with the earthquakes of the Mediterranean. As time went on it became clear that most damaging earthquakes were in fact not caused by volcanic activity.

A coherent global geological explanation of the majority of earthquakes is in terms of what is called plate tectonics<sup>(1-3)</sup>. The basic idea is that the Earth's outermost part (called the lithosphere) consists of several large and fairly stable rock slabs called plates. The ten largest plates are mapped in Figure 1-2. Each plate extends to a depth of about 80 kilometers.

Moving plates of the Earth's surface (see Figures 1-2 and 1-3) provide mechanisms for a great deal of the seismic activity of the world. Collisions between adjacent lithospheric plates, destruction of the slab-like plate as it descends or subducts into a dipping zone beneath island arcs (see Figure 1-4), and spreading along midoceanic ridges are all mechanisms that produce significant straining and fracturing of crustal Thus, the earthquakes in these rocks. tectonically active boundary regions are called plate-edge earthquakes. The very hazardous shallow earthquakes of Chile, Peru, the eastern Caribbean, Central America, Southern Mexico, California, Southern Alaska, the Aleutians the Kuriles, Japan, Taiwan, the Philippines, Indonesia, New Zealand, the Alpine-Caucasian-Himalayan belt are of plate-edge type.

As the mechanics of the lithospheric plates become better understood, long-term predictions of place and size may be possible for plate-edge earthquakes. For example, many plates spread toward the subduction zones at rates of from 2 to 5 centimeters (about one to two inches) per year. Therefore in active arcs like the Aleutian and Japanese Islands and subduction zones like Chile and western Mexico, knowledge of the history of large earthquake occurrence might flag areas that currently lag in earthquake activity.

Many large earthquakes are produced by slip along faults connecting the ends of offsets in the spreading oceanic ridges and the ends of island arcs or arc-ridge chains (see Figure 1-2). In these regions, plates slide past each other along what are called *transform faults*. Considerable work has been done on the estimation of strong ground motion parameters for the design of critical structures in earthquake-prone countries with either transform faults or ocean-plate subduction tectonics, such as Japan, Alaska, Chile and Mexico. The Himalaya, the Zagros and Alpine regions are examples of mountain ranges formed by continent-to-continent collisions. These collision zones are regions of high

present day seismic activity. The estimation of seismic hazard along continental collision margins at tectonic plates has not as yet received detailed attention.



*Figure 1-3.* Continued drift of the Indian plate towards Asian plate causes major Himalayan earthquakes. (From *The Collision Between India and Eurasia*, by Molnar and Tapponnier. Copyright 1977 by Scientific American, Inc. All rights reserved)

While a simple plate-tectonic theory is an important one for a general understanding of earthquakes and volcanoes, it does not explain all seismicity in detail, for within continental regions, away from boundaries, large devastating earthquakes sometimes occur. These *intraplate* earthquakes can be found on nearly every continent.



Figure 1-4. A sketch of the Earth's crust showing mid-oceanic ridges and active continental margin along a deep trench

One example of such earthquakes is the Dashte-e-Bayaz earthquake of August 31, 1968 in north-eastern Iran. In the United States, the most famous are the major earthquake series of 1811-1812 that occurred in the New Madrid area of Missouri, along the Mississippi River and the 1886 Charleston, South Carolina earthquake. One important group for example, which seems to bear no simple mechanical relation to the present plate edges, occurs in northern China.

Such major internal seismic activity indicates that lithospheric plates are not rigid or free of internal rupture. The occurrence of intraplate earthquakes makes the prediction of earthquake occurrence and size difficult in many regions where there is a significant seismic risk.

#### **1.3.2** Dilatancy in the Crustal Rocks

The crust of the continents is a rocky layer with average thickness of about 30 km but which can be as thick as 50 km under high mountain ranges. Under the ocean, the crustal thickness is no more than about 5 km.

At a depth in the crust of 5 kilometers or so, the lithostatic pressure (due to the weight of the overlying rocks) is already about equal to the strength of typical uncracked rock samples at the temperature ( $500^{\circ}$  C) and pressure appropriate for that depth. If no other factors entered, the shearing forces required to bring about sudden brittle failure and frictional slip along a crack would never be attained; rather, the rock would deform plastically. A way around this problem was the discovery that the presence of water provides a mechanism for sudden rupture by reduction of the effective friction along crack boundaries. Nevertheless, in a normal geological situation, such as the crust of coastal California, temperatures increase sufficiently fast so that at crustal depths greater than about 16 km the elastic rocks become viscoelastic. Strain is then relieved by slow flow or creep rather than by brittle fracture. The part of the crust above this transition point is the seismogenic zone.

Studies of the time of travel of P and S waves before the 1971 San Fernando earthquake indicated that four years before it occurred, the ratio of the velocity of the P waves to the velocity of the S waves decreased rather suddenly by 10 percent from its average value of 1.75. There was, thereafter, a steady increase in this ratio back to a more normal value. One explanation is the *dilatancy model*. This states that as the crustal rocks become strained, cracking occurs locally and the volume of rock increases or dilates. Cracking may occur too quickly for ground water to flow into the dilated volume to fill the spaces so the cracks become vapor-filled. The consequent fall in pore pressure leads to a reduction mainly in P wave velocities. Subsequent diffusion of ground water into the dry cracks increases the pore pressure, and provides water for lubrication along the walls of the cracks, while at the same time, the P wave velocity increases again (see Figure 1-31 in Section 1.10 below).

The full implications and relevance of the dilatancy theory of earthquake genesis are not yet clear, but the hypothesis is attractive in that it is consistent with precursory changes in ground levels, electrical conductivity and other physical properties which have been noted in the past before earthquakes. The theory has a potential for forecasting earthquakes under certain circumstances. For example, measurement of the P wave velocity in the vicinity of large reservoirs before and after

impounding of water might provide a more direct method of indicating an approaching seismic crisis near dams than is now available.

#### 1.3.3 Explosions

Ground shaking may be produced by the underground detonation of chemicals or nuclear devices. When a nuclear device is detonated in a borehole underground, enormous nuclear energy is released. Underground nuclear explosions fired during the past several decades at a number of test sites around the world have produced substantial artificial earthquakes (up to magnitude 6.0). Resultant seismic waves have traveled throughout the Earth's interior to be recorded at distant seismographic stations.

#### **1.3.4** Volcanic Earthquakes

As Figure 1-4 shows, volcanoes and earthquakes often occur together along the margins of plates around the world. Like earthquakes, there are also intraplate volcanic regions, such as the Hawaiian volcanoes.

Despite these tectonic connections between volcanoes and earthquakes, there is no evidence that moderate to major shallow earthquakes are not essentially all of tectonic, elastic-rebound type. Those earthquakes that can be reasonably associated with volcanoes are relatively rare and fall into three categories: (i) volcanic explosions, (ii) shallow earthquakes arising from magma movements, and (iii) sympathetic tectonic earthquakes.

Among the three categories, Category (iii), tectonically associated with volcanoes, is more difficult to tie down, as cases which may fit this category, are rare. There is no report of significantly increased volcanic activity in the great 1964 Alaska earthquake, but Puyehue Volcano in the Andes erupted 48 hours after the great 1960 Chilean earthquake. One might suppose that in a large earthquake the ground shaking would set up waves in reservoirs of magma; the general compression and dilatation of the gaseous liquid melt may trigger volcanic activity.

#### **1.3.5** Collapse Earthquakes

Collapse earthquakes are small earthquakes occurring in regions of underground caverns and mines. The immediate cause of ground shaking is the sudden collapse of the roof of the mine or cavern. An often observed variation is the *mine burst*. This rock rupture happens when the induced stress around the mine workings causes large masses of rock to fly off the mine face explosively, producing seismic waves. Mine bursts have been observed, for example, in Canada, and are especially common in South Africa.

An intriguing variety of collapse earthquakes is sometimes produced by massive landsliding. For example, a spectacular landslide on April 25, 1974, along the Mantaro River, Peru, produced seismic waves equivalent to a magnitude 4.5 earthquake. The slide had a volume of 1.6  $\times 10^9$  cubic meters and killed about 450 people.

#### 1.3.6 Large Reservoir-Induced Earthquakes

It is not a new idea that earthquakes might be triggered by impounding surface water. In the 1870's, the U.S. Corps of Engineers rejected proposals for major water storage in the Salton Sea in southern California on the grounds that such action might cause earthquakes. The first detailed evidence of such an effect came with the filling of Lake Mead behind Hoover Dam (height 221 meters), Nevada-Arizona, beginning in 1935. Although there may have been some local seismicity before 1935, after 1936 earthquakes were much more common. Nearby seismographs subsequently showed that after 1940, the seismicity declined. The foci of hundreds of detected earthquakes cluster on steeply dipping faults on the east side of the lake and have focal depths of less than 8 kilometers.

In Koyna, India, an earthquake (magnitude 6.5) centered close to the dam (height 103

caused significant damage meters) on December 11, 1967. After impounding began in 1962, reports of local shaking became prevalent in a previously almost aseismic area. that Seismographs showed foci were concentrated at shallow depths under the lake. In 1967 a number of sizable earthquakes occurred, leading up to the principal earthquake of magnitude 6.5 on December 11. This ground motion caused significant damage to buildings nearby, killed 177 persons, and injured more than 1,500. A strong motion seismograph in the dam gallery registered a maximum acceleration of 0.63g. The series of earthquakes recorded at Koyna has a pattern that seems to follow the rhythm of the rainfall (see Figure 1-5). At least a comparison of the frequency of earthquakes and water level suggests that seismicity increases a few months after each rainy season when the reservoir level is highest. Such correlations are not so clear in some other examples quoted.

In the ensuing years, similar case histories have been accumulated for several dozen large dams, but only a few are well documented. Most of these dams are more than 100 meters high and, although the geological framework at the sites varies, the most convincing examples of reservoir induced earthquakes occur in tectonic regions with at least some history of earthquakes. Indeed, most of the thousands of large dams around the world give no sign of earthquake induction. A poll in 1976 showed that only four percent of large dams had an earthquake reported with magnitude greater than 3.0 within 16 kilometers of the dam.

Calculation shows that the stress due to the load of the water in even large reservoirs is too small to fracture competent rock. The best explanation is that the rocks in the vicinity of the reservoir are already strained from the tectonic forces so that existing faults are almost ready to slip. The reservoir either adds a stress perturbation which triggers a slip or the increased water pressure lowers the strength of the fault, or both.



*Figure 1-5.* The relationship between reservoir level and local seismic activity at Koyna Dam. (From Earthquakes, by Bruce A. Bolt. Copyright 1978, 1999 W.H. Freeman and Company. Used with permission.)



*Figure 1-6.* Normal fault at the Corinth Canal, Greece. (Photo courtesy of L. Weiss.)

### **1.4** Earthquake fault sources

Field observations show that abrupt changes in the structure of rocks are common. In some places one type of rock can be seen butting up against rock of quite another type along a plane of contact. Such offsets of geological structure are called *faults*<sup>(1-4)</sup>. Clear vertical offset of layers of rock along an exposed fault in the wall of the Corinth canal, Greece, can be seen in Figure 1-6.

Faults may range in length from a few meters to many kilometers and are drawn on a geological map as continuous or broken lines (see Figure 1-7). The presence of such faults indicates that, at some time in the past, movement took place along them. Such movement could have been either *slow slip*, which produces no ground shaking, or *sudden rupture* (an earthquake). Figure 1-8 shows one of the most famous examples of sudden fault rupture slips of the San Andreas fault in April 1906. In contrast, the observed surface faulting



*Figure 1-7.* A simplified fault map of California. (From *The San Andreas Fault*, by Don L. Anderson. Copyright 1971 by Scientific American, Inc. Al rights reserved.)



*Figure 1-8.* Right-lateral horizontal movement of the San Andreas Fault in the 1906 earthquake the Old Sir Francis Highway. (Photo by G.K Gilbert, courtesy of USGS.)

of most shallow focus earthquakes is much shorter and shows much less offset. Indeed, in the majority of earthquakes, fault rupture does not reach the surface and consequently is not directly visible. Geological mappings and geophysical work show that faults seen at the surface sometimes extend to depths of tens of kilometers in the Earth's crust.

It must be emphasized that most faults plotted on geological maps are now inactive. However, sometimes previously unrecognized active faults are discovered from fresh ground breakage during an earthquake. Thus, a fault was delineated by a line of cracks in open fields south of Oroville after the Oroville dam, California earthquake of August 1, 1975. The last displacement to occur along a typical fault may have taken place tens of thousands or even millions of years ago. The local disruptive forces in the Earth nearby may have subsided long ago and chemical processes involving water movement may have healed the ruptures, particularly at depth. Such an inactive fault is not now the site of earthquakes and may never be again.

In seismology and earthquake engineering, the primary interest is of course in active faults, along which rock displacements can be expected to occur. Many of these faults are in well defined plate-edge regions of the Earth, such as the mid-oceanic ridges and young mountain ranges. However, sudden fault displacements can also occur away from regions of clear present tectonic activity (see Section 1.3.1).

Fault displacement in an earthquake may be almost entirely horizontal, as it was in the 1906 San Francisco earthquake along the San Andreas fault, but often large vertical motions occur, (Fig. 1-9) such as were evident in the 1992 Landers earthquake. In California in the 1971 San Fernando earthquake, an elevation change of three meters occurred across the ruptured fault in some places.



Figure 1-9. Normal fault scarp associated with the 1992 Landers, California, earthquake (Photo by Dr. Marshall Lew).





The classification of faults depends only on the geometry and direction of relative slip. Various types are sketched in Figure 1-10. The *dip* of a fault is the angle that fault surface makes with a horizontal plane and the *strike* is the direction of the fault line exposed at the ground surface relative to the north. A *strike-slip* fault, sometimes called a transcurrent fault, involves displacements of rock laterally, parallel to the strike. If when we stand on one side of a fault and see the motion on the other side is from left to right, the fault is *right-lateral* strike-slip. Similarly, we can identify *left-lateral* strike-slip.

A *dip-slip* fault is one in which the motion is largely parallel to the dip of the fault and thus has vertical components of displacement. A *normal* fault is one in which the rock above the inclined fault surface moves downward relative to the underlying crust. Faults with almost vertical slip are also included in this category.

A *reverse* fault is one in which the crust above the inclined fault surface moves upward relative to the block below the fault. *Thrust* faults are included in this category but are generally restricted to cases when the dip angle is small. In *blind thrust faults*, the slip surface does not penetrate to the ground surface.

In most cases, fault slip is a mixture of strike-slip and dip-slip and is called *oblique* faulting.

For over a decade it has been known that displacement in fault zones occurs not only by sudden rupture in an earthquake but also by slow differential slippage of the sides of the fault. The fault is said to be undergoing *tectonic creep*. Slippage rates range from a few millimeters to several centimeters.

The best examples of fault creep come from the San Andreas zone near Hollister, California, where a winery built straddling the fault trace is being slowly deformed; in the town, sidewalks, curbs, fences and homes are being offset. On the Hayward fault, on the east side of San Francisco Bay, many structures are being deformed and even seriously damaged by slow slip, including a large water supply tunnel, a drainage culvert and railroad tracks that intersect the zone.

Horizontal fault slippage has now also been detected on other faults around the world, including the north Anatolian fault at Ismetpasa in Turkey and along the Jordan Valley rift in Israel. Usually, such episodes of fault slip are *aseismic*-that is, they do not produce local earthquakes.

It is sometimes argued that a large damaging earthquake will not be generated along a fault that is undergoing slow fault slip, because the slippage allows the strain in the crustal rocks to be relieved periodically without sudden rupture. However, an alternative view is also plausible. Almost all fault zones contain a plastic clay-like material called *fault gouge*. It may be that, as the elastic crystalline rocks of the deeper crust stain elastically and accumulate the energy to be released in an earthquake, the weak gouge material at the top of the fault zone is carried along by the adjacent stronger rock to the side and underneath. This would mean that the slow slip in the gouge seen at the surface is an indication that strain is being stored in the basement rocks. The implication of this view is that, on portions of the fault where slippage occurs, an earthquake at depth could result from sudden rupture, but surface offset would be reduced. On the portion where slippage is small or nonexistent, offsets would be maximum. A prediction of this kind can be checked after earthquakes occur near places where slippage is known to be taking place.

Sometimes aseismic slip is observed at the ground surface along a ruptured fault that has produced an earlier substantial earthquake. For example, along the San Andreas fault break in the 1966 earthquake on June 27 near Parkfield, California, offset of road pavement increased by a few centimeters in the days following the main earthquake. Such continued adjustment of the crustal rock after the initial major offset is probably caused partly by aftershocks and partly by the yielding of the weaker surface rocks and gouge in the fault zone as they accommodate to the new tectonic pressures in the region.

It is clear that slow slippage, when it occurs in built up areas, may have unfortunate economic consequences. This is another reason why certain types of structures should not be built across faults if at all possible. When such structures including dams and embankments must be laid across active faults, they should have jointed or flexible sections in the fault zone.



*Figure 1-11.* Ground Motion near the ground surface due to P waves. (From *Nuclear Explosions and Earthquakes*, by Bruce A. Bolt. Copyright 1976 W. H. Freeman and Company. Used with Permission.)

### 1.5 seismic waves

Three basic types of elastic waves make up the shaking that is felt and causes damage in an earthquake<sup>(1-1)</sup>. These waves are similar in many important ways to the familiar waves in air, water, and gelatin. Of the three, only two propagate within a body of solid rock. The faster of these body waves is appropriately called the primary or *P wave*. Its motion is the same as that of a sound wave, in that, as it spreads out, it alternately pushes (compresses) and pulls (dilates) the rock (see Figure 1-11). These P waves, just like sound waves, are able to travel through both solid rock, such as granite mountains, and liquid material, such as volcanic magma or the water of the oceans.

The slower wave through the body of rock is called the secondary or *S wave*. As an S wave propagates, it shears the rocks sideways at right angles to the direction of travel (see Figure 1-12). Thus, at the ground surface S waves can produce both vertical and horizontal motions.

The S waves cannot propagate in the liquid parts of the Earth, such as the oceans and their amplitude is significantly reduced in liquefied soil.

The actual speed of P and S seismic waves depends on the density and elastic properties of the rocks and soil through which they pass. In most earthquakes, the P waves are felt first<sup>(1-5)</sup>. The effect is similar to a sonic boom that bumps and rattles windows. Some seconds later the S waves arrive with their significant component of side-to-side motion, so that the ground shaking is both vertical and horizontal. This S wave motion is most effective in damaging structures.

The speed of P and S waves is given in terms of the density of the elastic material and the elastic moduli. We let *k* be the modulus of incompressibility (bulk modulus) and  $\mu$  be the modulus of rigidity and  $\rho$  be the density. Then we have<sup>(1-5)</sup> for P waves;

$$V_{p} = \sqrt{\frac{k + \frac{4\mu}{3}}{\rho}}$$

for S waves;

$$V_{s} = \sqrt{\frac{\mu}{\rho}}$$
(1-2)

(1-1)

The third general type of earthquake wave is called a *surface wave* because its motion is restricted to near the ground surface. Such waves correspond to ripples of water that travel across a lake. Most of the wave motion is located at the outside surface itself, and as the depth below this surface increases, wave displacements become less and less.

Surface waves in earthquakes can be divided into two types. The first is called a Love wave. Its motion is essentially the same as that of S waves that have no vertical displacement; it moves the ground side to side in a horizontal plane parallel to the Earth's surface, but at right angles to the direction of propagation, as can be seen from the illustration in Figure 1-13. The second type of surface wave is known as a Rayleigh wave. Like rolling ocean waves, the pieces of rock disturbed by a Rayleigh wave move both vertically and horizontally in a vertical plane pointed in the direction in which the waves are travelling. As shown by the arrows in Figure 1-14. Each piece of rock moves in an ellipse as the wave passes.

Surface waves travel more slowly than body waves and, of the two surface waves, Love waves generally travel faster than Rayleigh waves. Thus, as the waves radiate outwards from the earthquake source into the rocks of the Earth's crust, the different types of waves separate out from one another in a predictable pattern.



*Figure 1-12.* Ground motion near the ground surface due to S waves. (From *Nuclear Explosions and Earthquakes*, by Bruce A. Bolt. Copyright 1976 W. H. Freeman and Company. Used with Permission.)

An illustration of the pattern seen at a distant place is shown in Figure 1-15. In this example, the seismograph has recorded only the vertical motion of the ground, and so the seismogram contains only P, S and Rayleigh waves, because Love waves are not recorded by vertical instruments.

When the body waves (the P and S waves) move through the layers of rock in the crust they are reflected or refracted at the interfaces between rock types, as illustrated in Figure 1-16a. Also, whenever either one is reflected or refracted, some of the energy of one type is converted to waves of the other type (see Figure 1-16b).



*Figure 1-13.* Ground motion near the ground surface due to Love waves. (From *Nuclear Explosions and Earthquakes*, by Bruce A. Bolt. Copyright 1976 W. H. Freeman and Company. Used with Permission.)



*Figure 1-15.* A seismograph record of the vertical component of a distant earthquake (third trace on bottom) on which the arrival of P, S and Rayleigh waves are marked. (Time increases from left to right.)



*Figure 1-14.* Ground motion near the ground surface due to Rayleigh waves. (From *Nuclear Explosions and Earthquakes*, by Bruce A. Bolt. Copyright 1976 W. H. Freeman and Company. Used with Permission.)

When P and S waves reach the surface of the ground, most of their energy is reflected back into the crust, so that the surface is affected almost simultaneously by upward and downward moving waves. For this reason *considerable amplification of shaking typically occurs near the surface*-sometimes doubling the amplitude of the upcoming waves.

This surface amplification enhances the shaking damage produced at the surface of the Earth. Indeed, in many earthquakes mine workers below ground report less shaking than people on the surface.





*Figure 1-16.* Reflection, refraction, and transformation of body waves. (From *Nuclear Explosions and Earthquakes*, by Bruce A. Bolt. Copyright 1999 W. H. Freeman and Company. Used with permission.)

Date	Region	Deaths	Magnitude (M <sub>S</sub> )
December 7, 1988	Spitak, Armenia	25,000	7.0
August 1, 1989	West Iran, Kurima District	90	5.8
October 17, 1989	Santa Cruz Mountains, Loma Prieta	63	7.0
June 20, 1990	Caspian Sea, Iran	Above 40,000	7.3
March 13, 1992	Erzinean, Turkey	540	6.8
July 16, 1990	Luzon, Phillipines	1,700	7.8
July 12, 1993	Hokkaido, Japan	196	7.8
September 29, 1993	Killari, India	10,000	6.4
January 17, 1994	Northridge, California	61	6.8
January 16, 1995	Kobe, Japan	5400	6.9
August 17, 1999	Izmit, Turkey	16,000	7.4
September 21, 1999	Chi Chi, Taiwan	2,200	7.6

Table 1-1. Magnitudes of Some Recent Damaging Earthquakes

Another reason for modification of the incoming seismic wave amplitudes near the ground surface is the effect of layers of weathered rock and soil. When the elastic moduli have a mismatch from one layer to another, the layers act as wave filters amplifying the waves at some frequencies and deamplifying them at others. Resonance effects at certain frequencies occur.

Seismic waves of all types are progressively damped as they travel because of the nonelastic properties of the rocks and soils. The attenuation of S waves is greater than that of P waves, but for both types attenuation increases as wave frequency increases. One useful seismological quantity to measure damping is the parameter Q such that the amplitude A at a distance d of a wave frequency f (Hertz) and velocity C is given by:

$$A = A_0 e^{-\left(\pi f d / QC\right)} \tag{1-3}$$

For P and S waves in sediments, Q is about 500 and 200, respectively.

The above physical description is approximate and while it has been verified closely for waves recorded by seismographs at a considerable distance from the wave source (*the far-field*), it is not adequate to explain important details of the heavy shaking near the center of a large earthquake (the near-field). Near a fault that is suddenly rupturing, the strong ground shaking in the associated earthquake consists of a mixture of various kinds of seismic waves that have not separated very distinctly. To complicate the matter, because the source of radiating seismic energy is itself spread out across an area, the type of ground motion may be further mixed together. This complication makes identification of P, S and surface waves on strong motion records obtained near to the rupturing fault particularly difficult. However, much progress in this skill, based on intense study and theoretical modeling, has been made in recent years. This advance has made possible the computation of realistic ground motions at specified sites for engineering design purposes<sup>(1-6)</sup>.

A final point about seismic waves is worth emphasizing here. There is considerable evidence, observational and theoretical, that earthquake waves are affected by both soil conditions and topography. For example, in weathered surface rocks, in alluvium and water-filled soil, the size of P, S and surface waves may be either increased or decreased depending on wave frequency as they pass to and along the surface from the more rigid basement rock.



Figure 1-17. Tilting of buildings due to soil liquefaction during the Niigata (Japan) earthquake of 1964

The wave patterns from earthquake sources are much affected by the three-dimensional nature of the geological structures<sup>(1-6)</sup>. Clear evidence on this effect comes from recordings of the 1989 Loma Prieta earthquake (see Table 1-1). First, strong motion recordings show that there were reflections of high frequency S waves from the base of the Earth's crust at a depth of about 20 km, under southern San Francisco Bay. Secondly, the seismic S waves, like light waves, are polarized by horizontal layering into a horizontal component (SH type) and a vertical component (SV type). Because of large differences in the rock structure from one side of the San Andreas fault to the other, there were also lateral variations by refraction of SH waves across this deep crustal velocity contrast. This produced significant amplitude variations, with azimuth from the seismic source, of the strong ground shaking in a period range of about 1 to 2 seconds. In addition, there was measurable scattering of shear waves by deep alluvial basins south of San Francisco Bay. In sum, the large wave amplitudes caused enhanced intensity in a region between San Francisco and Oakland, about 10 km wide by 15 km long.

Finally, it should be noted that seismic S waves travel through the rocks and soils of the Earth with a rotational component. Torsional components of ground motion are thought to have important effects on the response of certain types of structures. Some building codes now contain material on practices that take rotational ground motion into consideration.

# 1.6 EARTHQUAKE DAMAGE MECHANISMS

Earthquakes can damage structures in various ways such as:

- 1. by inertial forces generated by severe ground shaking.
- 2. by earthquake induced fires.
- 3. by changes in the physical properties of the foundation soils (e.g. consolidation, settling, and liquefaction).
- 4. by direct fault displacement at the site of a structure.
- 5. by landslides, or other surficial movements.
- by seismically induced water waves such as seismic sea waves (tsunamis) or fluid motions in reservoirs and lakes (*seiches*).
- 7. by large-scale tectonic changes in ground elevation.

Of the above categories, by far the most serious and widespread earthquake damage and accompanying loss of life are caused by severe ground shaking. The bulk of this handbook is devoted to design techniques and measures for reducing this type of hazard<sup>(1-7)</sup>.

Fire hazards in earthquakes must also be emphasized. Vivid memories remain of the great conflagrations that followed the San Francisco 1906 earthquake and Tokyo's 1923 earthquake. In the 1906 San Francisco earthquake perhaps 20 percent of the total loss was due directly to ground motions. However, the fire, which in three days burned 12 square kilometers and 521 blocks of downtown San Francisco, was the major property hazard.

Soil related problems have caused major economic loss in past earthquakes. One classic example of this type of damage happened in the 1964 earthquake of Niigata, Japan. The maximum ground acceleration was



Figure 1-18. Faults rupturing under Managua (Nicaragua) during the earthquake of 1972

approximately 0.16g which considering the amount of damage, is not high. Expansion of the modern city of Niigata had involved reclamation of land along the Shinano River. In the newly deposited and reclaimed land areas many buildings tilted or subsided as a result of soil liquefaction (see Figure 1-17). 3,018 houses were destroyed and 9,750 were moderately or severely damaged in Niigata prefecture alone, most of the damage was caused by cracking and unequal settlement of the ground. About 15,000 houses in Niigata city were inundated by the collapse of a protective embankment along the Shinano River. The number of deaths was only 26. Precautionary and design measures against earthquake induced soil problems are discussed in Chapter 3.

Perhaps surface fault displacements are the most frightening aspect of earthquakes to the general public. However, although severe local damage has occurred in this way, compared with damage caused by strong ground shaking, this type of damage is rather rare. Even in very large earthquakes, the area exposed to direct surface fault displacement is much smaller than the area affected by strong ground shaking. One of the clearest examples of damage caused by direct fault displacement occurred in the Managua, Nicaragua earthquake of 1972, where four distinct faults ruptured under the city (see Figure 1-18). The total length of fault rupture within the city was about 20 kilometers, and the maximum fault displacement on two of the faults reached about 30 centimeters. Even in this case the total area damaged by direct faulting was less than one percent of the area damaged by strong ground shaking.

Earthquake-induced landslides and avalanches, although responsible for major devastation, are fortunately localized. The most pronounced example of this kind of damage occurred in Peru earthquake of May 31, 1970. This magnitude 7.75 earthquake led to the greatest seismological disaster yet experienced in the Western Hemisphere. An enormous debris avalanche from the north peak of Huascaran Mountain (see Figure 1-19) amounting to 50,000,000 or more cubic meters of rock, snow, ice, and soil, travelled 15 kilometers from the mountain to the town of Yungay with an estimated speed of 320 kilometers per hour. At least 18,000 people were buried under this avalanche, which covered the towns of Ranrahirca and most of Yungay.

Earthquake-induced changes in ground elevations (see Figure 1-20) may not cause major injuries or loss of life. Their most important threat is the damage they can cause to structures such as bridges and dams.

Seismic sea waves, or *tsunamis*, are long water waves generated by sudden displacements under water. The most common cause of significant tsunamis is the impulsive displacement along a submerged fault, associated with a large earthquake. Because of the great earthquakes that occur around the Pacific, this ocean is particularly prone to seismic sea waves.



*Figure 1-19.* Aerial view of Mt. Huascaran and the debris avalanche that destroyed Yungay and Ranrahirca in May 1970 Peru earthquake. (Photo courtesy of Servicio Aerofotografico National de Peru and L. Cluff.)

For earthquakes to generate tsunamis, dipslip faulting (see Figure 1-10) seems to be necessary, and strike-slip faulting is almost never accompanied by damaging tsunamis. History contains many accounts of great offshore earthquakes being accompanied by destructive tsunamis. In June 15, 1896, in Honshu region of Japan a tsunami with a visual run-up height exceeding 20 meters (65 feet) drowned about 26,000 people. More recently, the Chilean earthquake of 1960 caused a tsunami with a run-up height of 10 meters at Hilo, Hawaii (see Figure 1-21). A tsunami at Crescent City, California, caused by the great Alaskan earthquake of 1964, resulted in 119 deaths and over \$104,000,000 damage.

The most important scheme to prevent loss of life in the Pacific from tsunamis is the Seismic Sea Wave Warning System. The warning system is made up of a number of seismological observatories including Berkeley, California; Tokyo, Japan; Victoria, Canada and about 30 tide stations around the Pacific Ocean. The time of travel of a tsunami wave from Chile to the Hawaiian islands is, for example, about 10 hours and from Chile to Japan about 20 hours. Under this system, therefore, there is ample time for alerts to be followed up by local police action along coastlines so that people can be evacuated.

Apart from the tsunami warning system, the hazard can be mitigated by using adequate design of wharf, breakwater and other facilities based on techniques of coastal engineering. Often, however, zoning around coastlines is desirable to prevent building in the most lowlying areas where tsunamis are known to overwash the surface level. Sufficient information is nowadays usually available to allow local planners to make prudent decisions.



Figure. 1-20 Ground uplift along the fault in the 1999 Chi-Chi Earthquake (Photo by Dr. Farzad Naeim).



Before tsunami



After tsunami

Figure 1-21 Damage at Hilo, Hawaii, due to tsunami of May 23, 1960. (Photos courtesy of R. L. Wiegel.)

## 1.7 QUANTIFICATION OF EARTHQUAKES

#### **1.7.1** Earthquake Intensity

The oldest useful yardstick of the "strength" of an earthquake is earthquake *intensity*<sup>(1-1)</sup>. Intensity is the measure of damage to works of man, to the ground surface, and of human reaction to the shaking. Because earthquake intensity assessments do not depend on instruments, but on the actual observation of effects in the meizoseismal zone, intensities can

be assigned even to historical earthquakes. In this way, the historical record becomes of utmost importance in modern estimates of seismological risk.

The first intensity scale was developed by de Rossi of Italy and Forel of Switzerland in the 1880s. This scale, with values from I to X, was used for reports of the intensity of the 1906 San Francisco earthquake, for example. A more refined scale was devised in 1902 by the Italian

volcanologist and seismologist Mercalli with a twelve-degree range from I to XII. A version is given in Table 1-2, as modified by H.O. Wood

#### Table 1-2. Modified Mercalli Intensity Scale (MMI) of 1931

- I. Not felt except by a very few under especially favorable circumstances.
- II. Felt only by a few persons at rest, especially on upper floors of buildings. Delicately suspended objects may swing.
- III. Felt quite noticeably indoors, especially on upper floors or buildings, but many people do not recognize it as an earthquake. Standing motor cars may rock slightly. Vibration like passing of truck. Duration estimated.
- IV. During the day felt indoors by many, outdoors by few. At night some awakened. Dishes, windows, doors disturbed; walls make cracking sound. Sensation like heavy truck striking building. Standing motor cars rocked noticeably.
- V. Felt by nearly everyone, many awakened. Some dishes, windows, etc., broken; a few instances of cracked plaster; unstable objects overturned. Disturbances of trees, poles, and other tall objects sometimes noticed. Pendulum clocks may stop.
- VI. Felt by all, many frightened and run outdoors. Some heavy furniture moved; a few instances of fallen plaster or damaged chimneys. Damage slight.
- VII. Everybody runs outdoors. Damage negligible in buildings of good design and construction; slight to moderate in well-built ordinary structures; considerable in poorly built or badly designed structures; some chimneys broken. Noticed by persons driving motor cars.
- VIII. Damage slight in specially designed structures; considerable in ordinary substantial buildings, with partial collapse; great in poorly built structures. Panel walls thrown out of frame structures. Fall of chimneys, factory stacks, columns, monuments, walls. Heavy furniture overturned. San and mud ejected in small amounts. Changes in well water. Persons driving motor cars disturbed.
- IX. Damage considerable in specially designed structures; well designed frame structures thrown out of plumb; great in substantial buildings, with partial collapse. Buildings shifted off foundations. Ground cracked conspicuously. Underground pipes broken.
- X. Some well-built wooden structures destroyed; most masonry and frame structures destroyed with foundations; ground badly cracked. Rails bent. Landslides considerable from river banks and steep sloes. Shifted sand and mud. Water splashed (slopped) over banks.
- XI. Few, if any, (masonry) structures remain standing. Bridges destroyed. Broad fissures in ground. Underground pipelines completely out of service. Earth slumps and land slips in soft ground. Rail bent greatly.
- XII. Damage total. Practically all works of construction are damaged greatly or destroyed. Waves seen on ground surface. Lines of sight and level are distorted. Objects are thrown into the air.

#### Table 1-3. Japanese Seismic Intensity Scale

- 0. Not felt; too weak to be felt by humans; registered only by seismographs.
  - I. Slight: felt only feebly by persons at rest or by those who are sensitive to an earthquake.
  - II. Weak: felt by most persons, causing light shaking of windows and Japanese latticed sliding doors (shoji).
- III. Rather strong: shaking houses and buildings, heavy rattling of windows and Japanese latticed sliding doors, swinging of hanging objects, sometimes stopping pendulum clocks, and moving of liquids in vessels. Some persons are so frightened as to run out of doors.
- IV. Strong: resulting in strong shaking of houses and buildings. Overturning of unstable objects, spilling of liquid out of vessels.
- V. Very strong: causing cracks in brick and plaster walls, overturning of stone lanterns and grave stones, etc. and damaging of chimneys and mud and plaster warehouses. Landslides in steep mountains are observed.
- VI. Disastrous: causing demolition of more that 1% of Japanese wooden houses; landslides, fissures on flat ground accompanied sometimes by spouting of mud and water in low fields.
- VII. Ruinous: causing demolition of almost all houses: large fissures and faults are observed.

and Frank Neumann to fit conditions in California<sup>(1-7)</sup>. The descriptions in Table 1-2 allow the damage to places affected by an earthquake to be rated numerically. These spot intensity ratings can often be separated by lines which form an isoseismal map (see Figure 1-22). Such intensity maps provide crude, but valuable information on the distribution of strong ground shaking, on the effect of surficial soil and underlying geological strata, the extent of the source, and other matters pertinent to insurance and engineering problems.

Because intensity scales are subjective and depend upon social and construction conditions of a country, they need revising from time to time. Regional effects must be accounted for. In this respect, it is interesting to compare the Japanese scale (0 to VII) summarized in Table 1-3 with the Modified Mercalli descriptions.

#### **1.7.2** Earthquake Magnitude

If sizes of earthquakes are to be compared world-wide, a measure is needed that does not depend, as does intensity, on the density of population and type of construction. A strictly quantitative scale that can be applied to earthquakes in both inhabited and uninhabited regions was originated in 1931 by Wadati in Japan and developed by Charles Richter in 1935 in California.



*Figure 1-22.* A typical isoseismal map. Similar maps are now plotted by the *Trinet* program in Southern California.

Richter<sup>(1-5)</sup> defined the magnitude of a local earthquake as the logarithm to base ten of the maximum seismic wave amplitude in microns (10<sup>-4</sup> centimeters) recorded on a Wood-Anderson seismograph located at a distance of 100 kilometers from the earthquake epicenter (see Figure 1-23). This means that every time the magnitude goes up by one unit, the amplitude of the earthquakes waves increase 10 times. Since the fundamental period of the Wood-Anderson seismograph is 0.8 second, it selectively amplifies those seismic waves with a period ranging approximately from 0.5 to 1.5 seconds. Because the natural period of many building structures are within this range, the local Richter magnitude remains of value to engineers.

It follows from the definition of the magnitude, that it has no theoretical upper or lower limits. However, the size of an earthquake is limited at the upper end by the strength of the rocks of the Earth's crust. Since 1935, only a few earthquakes have been recorded on seismographs that have had a magnitude over 8.0. At the other extreme, highly sensitive seismographs can record earthquakes with a magnitude of less than minus two. See Table 1-4 for an average number of world-wide earthquakes of various magnitudes.

Generally speaking, shallow earthquakes have to attain Richter magnitudes of more than 5.5 before significant widespread damage occurs near the source of the waves.

At its inception, the idea behind the Richter local magnitude scale  $(M_L)$  was a modest one. It was defined for Southern California, shallow earthquakes, and epicentral distances less than about 600 kilometers. Today, the method has been extended to apply to a number of types of seismographs throughout the world (see Figure 1-24). Consequently a variety of magnitude scales based on different formulas for epicentral distance and ways of choosing an appropriate wave amplitude, have emerged:

Surface Wave Magnitude  $(M_s)$  Surface waves with a period around 20 seconds are often dominant on the seismograph records of distant earthquakes (epicentral distances of more than 2000 kilometers). To quantify these earthquakes, Gutenberg defined a magnitude scale  $(M_s)$  which is based on measuring the amplitude of surface waves with a period of 20 seconds <sup>(1-8)</sup>.



Figure 1-23. Definition of local Richter magnitude

**Body Wave Magnitude**  $(m_b)$  Deep focus earthquakes have only small or insignificant trains of surface waves. Hence, it has become routine in seismology to measure the amplitude of the P wave, which is not affected by the focal depth of the source, and thereby determine a P wave magnitude  $(m_b)$ . This magnitude type has also been found useful in continental regions like the eastern United States where no Wood-Anderson instruments have operated historically.

**Moment Magnitude**  $(M_W)$  Because of significant shortcomings of  $M_L$ ,  $m_b$ , and to a lesser degree  $M_s$  in distinguishing between great earthquakes, the moment magnitude scale was devised<sup>(1-10)</sup>. This scale assigns a magnitude to the earthquake in accordance with its seismic moment  $(M_0)$  which is directly related to the size of the earthquake source:

$$M_W = (\log M_0) / 1.5 - 10.7 \tag{1-4}$$

where  $M_0$  is seismic moment in dyn-cm.

**Magnitude Saturation** As described earlier, the Richter magnitude scale  $(M_L)$  measures the seismic waves in a period range of particular importance to structural engineers (about 0.5-1.5 seconds). This range corresponds approximately to wave-lengths of 500 meters to 2 kilometers. Hence, although theoretically there is no upper bound to Richter magnitude, progressively it underestimates more seriously the strength of earthquakes produced by the longer fault rupture lengths. The saturation point for the Richter magnitude scale is about  $M_L = 7$ . The body wave magnitude  $(m_b)$  saturates at about the same value.



*Figure 1-24.* Amplification of seismic waves by various seismographs. (From *The Motion of Ground in Earthquakes*, by David M. Boore. Copyright 1977 by Scientific American, Inc. All rights reserved.)

Table 1-4. World-wide Earthquakes per Year		
Magnitude $M_s$	Average No. > $M_s$	
8	1	
7	20	
6	200	
5	3,000	
4	15,000	
3	> 100,000	

In contrast, the surface-wave magnitude  $(M_s)$  which uses the amplitude of 20 second surface waves (wave-length of about 60 kilometers) saturates at about  $M_s = 8$ . Its inadequacy in measuring the size of great earthquakes can be illustrated by comparing the San Francisco earthquake of 1906 and the Chilean earthquake of 1960 (see Figure 1-25). Both earthquakes had a magnitude  $(M_s)$  of 8.3.

However, the area that ruptured in the San Francisco earthquake (the dashed area) was approximately 15 kilometers deep and 400 kilometers long whereas the area that ruptured in the Chilean earthquake (the dotted area) was equal to about half of the state of California. Clearly the Chilean earthquake was a much larger event.

The moment-magnitude scale  $(M_W)$  is the only magnitude scale which does not suffer from the above mentioned saturation problem for great earthquakes. The reason is that it is directly based on the forces that work at the fault rupture to produce the earthquake and not the recorded amplitude of specific types of seismic waves. Hence, as can be expected, when moment magnitudes were assigned to the San Francisco earthquake of 1906 and the Chilean earthquake of 1960, the magnitude of the San Francisco earthquake dropped to 7.9, whereas the magnitude of the Chilean earthquake was raised to 9.5.  $M_s$  and  $M_w$  for some great earthquakes are compared in Table 1-5. Magnitudes of some recent damaging earthquakes are shown in Table 1-1.

In light of the above discussion, application of different scales have been suggested for measuring shallow earthquakes of various magnitudes:

$M_D$	for magnitudes less than 3
$M_L$ or $m_b$	for magnitudes between 3 and 7
$M_{s}^{-}$	for magnitudes between 5 and 7.5
$M_W$	for all magnitudes

Table 1-5. Magnitudes of Some of the Great

Earthquakes			
Date	Region	$M_s$	$M_w$
January 9, 1905	Mongolia	81⁄4	8.4
Jan. 31, 1906	Ecuador	8.6	8.8
April 18, 1906	San Francisco	81⁄4	7.9
Jan. 3, 1911	Turkestan	8.4	7.7
Dec. 16, 1920	Kansu, China	8.5	7.8
Sept. 1, 1923	Kanto, Japan	8.2	7.9
March 2, 1933	Sanrika	8.5	8.4
May 24, 1940	Peru	8.0	8.2
April 6, 1943	Chile	7.9	8.2
Aug. 15, 1950	Assam	8.6	8.6

Date	Region	$M_s$	$M_w$
Nov. 4, 1952	Kamchatka	8	9.0
March 9, 1957	Aleutian Islands	8	9.1
Nov. 6, 1958	Kurile Islands	8.7	8.3
May 22, 1960	Chile	8.3	9.5
March 28, 1964	Alaska	8.4	9.2
Oct. 17, 1966	Peru	7.5	8.1
Aug. 11, 1969	Kurile Islands	7.8	8.2
Oct, 3, 1974	Peru	7.6	8.1
July 27, 1976	China	8.0	7.5
Aug. 16, 1976	Mindanao	8.2	8.1
March 3, 1985	Chile	7.8	7.5
Sep. 19, 1985	Mexico	8.1	8.0

# 1.8 EARTHQUAKE SOURCE MODELS

Field evidence in the 1906 California earthquake showed clearly that the strained rocks immediately west of the San Andreas fault had moved north-west relative to the rocks to the east. Displacements of adjacent points along the fault reached a maximum of 6 meters near Olema in the Point Reyes region.

H.F. Reid <sup>(1-11)</sup> studied the triangulation surveys made by the U.S. Coast and Geodetic Survey across the region traversed by the 1906 fault break. These surveys made in 1851-1865, 1874-1892 and just after the earthquake, showed (i) small inconsistent changes in elevation along the San Andreas fault; (ii) significant horizontal displacements parallel to the fault trace; and (iii) movement of distant points on opposite sides of the fault of 3.2 meters over the 50-year period, the west side moving north.

Based on geological evidence, geodetic surveys, and his own laboratory experiments, Reid put forth the elastic rebound theory for source mechanism that would generate seismic waves. This theory supposes that the crust of the Earth in many places is being slowly displaced by underlying forces. Differential displacements set up elastic strains that reach



*Figure 1-25.* Fault rupture area for the San Francisco 1906 and Chile 1960 earthquakes. (Modified from *The Motion of Ground in Earthquakes*, By David M. Boore. Copyright 1977 by Scientific American, Inc. All rights reserved.)

levels greater than can be endured by the rock. Ruptures (faults) then occur, and the strained rock rebounds along the fault under the elastic stresses until the strain is partly or wholly relieved (see Figures 1-26 and 1-27). This theory of earthquake mechanism has been verified under many circumstances and has required only minor modification.



*Figure 1-26.* A bird's eye view of market lines drawn along a road AB, which crosses a fault trace at the ground surface. (a) Elastic strain accumulation before fault rupture. (b) Final position after the fault rupture. (From *Earthquakes*, by Bruce A. Bolt. Copyright 1999, W.H. Freeman and Company. Used with permission.)

The strain slowly accumulating in the crust builds a reservoir of elastic energy, in the same way as a coiled spring, so that at some place, the *focus*, within the strained zone, rupture suddenly commences, and spreads in all directions along the fault surface in a series of erratic movements due to the uneven strength of the rocks along the tear. This uneven propagation of the dislocation leads to bursts of high-frequency waves, which travel into the Earth to produce the seismic shaking that causes the damage to buildings. The fault rupture moves with a typical velocity of two to three kilometers per second and the irregular steps of rupture occur in fractions of a second. Ground shaking away from the fault consists of all types of wave vibrations with different frequencies and amplitudes.

In 1966, N. Haskell<sup>(1-12)</sup> developed a model "in which the fault displacement is represented by a coherent wave only over segments of the fault and the radiations from adjacent sections are assumed to be statistically independent or incoherent." The physical situation in this model is that the rupture begins suddenly and then spreads with periods of acceleration and retardation along the weakly welded fault zone. In this model, the idea of statistical randomness of fault slip or *chattering* in irregular steps along the fault plane is introduced.

More recently, Das and Aki<sup>(1-13)</sup> have considered a fault plane having various *barriers* distributed over it. They conceive that rupture would start near one of the barriers and then propagate over the fault plane until it is brought to rest or slowed at the next barrier. Sometimes the barriers are broken by the dislocation; sometimes the barriers remain unbroken but the dislocation reinitiates on the far side and continues; sometimes the barrier is not broken initially but, due to local repartitioning of the stresses and possible nonlinear effects, it eventually breaks, perhaps with the occurrence of aftershocks.

The elastic rebound model involving a moving dislocation along a fault plane segmented by barriers, over which roughnesses (or *asperities*) of various types are distributed stochastically, is thus the starting point for the modern interpretation of near-field records <sup>(1-14)</sup>. Based on this model, there have been recently a number of attempts to compute synthetic seismograms from points near to the source and comparisons have been made with observations (see Section 1.10).

As mentioned earlier, there are different kinds of fault ruptures. Some involve purely



Figure 1-27. Elastic rebound model of earthquakes

horizontal slip (strike-slip); some involve vertical slip (dip-slip). It might be expected that the wave patterns generated by fault geometries and mechanisms of different kinds will be different to a larger or lesser extent, because of the different radiation patterns produced. These geometries be modeled different can mathematically bv appropriate radiation functions.

The theory must also incorporate effects of source. The Doppler-like moving the consequences will depend on the speed of fault rupture and the directions of faulting. The physical problem is analogous (but more difficult) to the problem of sound emission from moving sources. The problem can be kinematically approached both and dynamically. The acoustic problem shows that in the far-field the pressure is the same as when

the source is at rest. However, in the near-field, the time dependence of both frequency and wave amplitude is a function of the azimuth of the site relative to the moving source (Figure 1-28).

In the case of a fault rupture toward a site at a more or less constant velocity (almost as large as the sear wave velocity), most of the seismic energy from the elastic rebound of the fault arrives in a single large pulse of motion (velocity or displacement), which occurs near the beginning of the record. This wave pulse represents the cumulative effect of almost all of the seismic radiation from the moving dislocation. In addition, the radiation pattern of the shear dislocation causes this large pulse of motion to be oriented mostly in the direction perpendicular to the fault. Coincidence of the radiation pattern maximum for tangential motion and the wave focusing due to the rupture propagation direction toward the site produces a large displacement pulse normal to the fault strike.

The horizontal recordings of stations in the 1966 Parkfield, California and the Pacoima station in the 1971 San Fernando, California earthquake were the first to be discussed in the literature as showing characteristic velocity pulses. These cases, with maximum amplitudes of 78 and 113 cm/sec, respectively, consisted predominantly of horizontally polarized SH wave motion and were relatively long period (about 2-3 sec). The observed pulses are consistent with the elastic rebound theory of earthquake genesis propounded by H.F. Reid after the 1906 San Francisco earthquake. These velocity and displacement pulses in the horizontal direction near the source were first called the source *fling*. Additional recordings in the near field of large sources have confirmed the presence of energetic pulses of this type, and they are now included routinely in synthetic ground motions for seismic design purposes. Most recently, the availability of instrumented measured ground motion close to the sources of the 1994 Northridge earthquake, the 1995 Kobe earthquake and the 1999 Chi-Chi earthquake provided important recordings of the "fling" or velocity pulse.

As in acoustics, the amplitude and frequency of the velocity and displacement pulses or "fling" have a geometrical focusing factor, which depends on the angle between the direction of wave propagation form the source and the direction of the source velocity. Instrumental measurements show that such directivity focusing can modify the amplitude velocity pulses by a factor of up to 10. The pulse may be single or multiple, with variations in the impetus nature of its onset and in its halfwidth period. A widely accepted illustration is the recorded ground displacement of the October 15, 1979 Imperial Valley, California, earthquake generated by a strike-slip fault source. The main rupture front moved toward El Centro and away from Bonds Corner.

We now summarize the main lines of



Figure 1-28. Effect of direction of fault rupture on ground motion experienced at a site. [After Benioff<sup>(1-15)</sup> and Singh <sup>(1-16)</sup>]

approach to modeling mathematically the earthquake source. The first model is the kinematic approach in which the time history of the slip on the generating fault is given a priori. Several defining parameters may be specified, such as the shape, duration, and amplitude of the source (or source time function and slip), the velocity of the slip over the fault surface, and the final area of the region over which the slip occurred. Numerous theoretical papers using this approach have been published (see the various discussions in Reference 1-16). The process is a kind of complicated curve fitting whereby the parameters of the source are varied in order to estimate by inspection the closeness of fit between recorded and computed nearfield or far-field seismic waves. Once the seismic source is defined by this comparison process, then the estimated source parameters can be used to extrapolate from the known ground motions near to a historical source to the future conditions required for engineering purposes.

A second approach is to use the differential equations involving the forces which produce the rupture. This dynamic procedure has received considerable emphasis lately. The basic model is a shear crack which is initiated in the pre-existing stress field and which causes stress concentrations around the tip of the crack. These concentrations, in turn, cause the crack to grow. For example, analytic expressions for particle accelerations in given directions from a uniformly growing elliptical crack are derived, but the effect of crack stoppage is not always included (this unrealistic boundary condition is included in most work of this kind).

The key to the crack problem seems to be in modeling the physical processes of the typical crack where there is interaction between the stress accumulation, rate of crack growth, and the criterion of fracture. Most studies on dynamic shear cracks are concerned primarily with the actual rupture process, and so the crack is assumed to be embedded in an infinite homogeneous medium. More realistic studies concerned with the seismic waves that are recorded in the near-field need a numerical approach, such as finite elements or finite differences, to handle geologic structural conditions.



Figure1-29. Definition of seismic moment.

The studies mentioned under kinematic and dynamic models are built around the elastic rebound theory of slip on a fault. There are, however, more general studies that take a less specific view of the earthquake source<sup>(1-16)</sup>.

It should be mentioned here that the scalar seismic moment (direction of force couples along the fault ignored) is given by

$$M_0 = \mu A D \tag{1-5}$$

where  $\mu$  is the rigidity of the material surrounding the fault, A is the slipped area, and D is the amount of slip (see Figure 1-29). The seismic moment is now the preferred parameter to specify quantitatively the overall size of an earthquake source.

Let us now summarize the physical model for the earthquake source generally accepted at present (see Figure 1-30). The source extends over a fault plane in the strained crustal rocks by a series of dislocations, which initiate at the focus and spread out with various rupture velocities. The dislocation front changes speed as it passes through patches on roughness (asperities on the fault).



*Figure1-30.* Simplified model of the vertical rupture surface for the Coyote Lake earthquake of 1979 in California

At the dislocation itself, there is a finite time for a slip to take place and the form of the slip is an elastic rebound of each side of the fault leading to a decrease of overall strain. The slip can have vertical components, as well as horizontal components, and can vary along the fault. The waves are produced near the dislocation front as a result of the release of the strain energy in the slippage<sup>(1-17)</sup>.

This model resembles in many ways radio waves being radiated from a finite antenna. In the far-field, the theory of radio propagation gives complete solutions for the reception of radio signals through stratified media. However, when the receiver is very near to the extended antenna, the signal becomes mixed because of the finiteness of the source and interference through end effects. The main parameters in the model are:

Rupture length	L
Rupture width	W
Fault slippage (offset)	D
Rupture velocity	V
Rise time	Т

Roughness (barrier) distribution density  $\phi(x)$ 

The main work in theoretical seismology on source properties today is to determine which of these parameters are essential, whether the set is an optimal one, and how best to estimate each parameter from both field observations and from analysis of the seismograms made in the near and the far-field.

A number of papers have now been published that demonstrate that, in certain important cases, synthetic seismograms for seismic waves near the source can be computed realistically<sup>(1-18)</sup>. The synthetic motions can be compared with the three observed orthogonal components of either acceleration, velocity, or displacement at a site. There remain difficulties, however, in modeling certain observed complexities and there is a lack of uniqueness in the physical formulations which lead to acceptable fits with observations (see also Section 1.10).

## 1.9 SEISMIC RISK EVALUATION

Regional seismicity or risk maps recommended by seismic design codes (see Chapter 5) usually do not attempt to reflect geological conditions nor to take into account variations due to soil properties. It is necessary, therefore, for critical construction in populated regions to make special geological-engineering studies for each site, the detail and level of concern which is used depending on the density of occupancy as well as the proposed structural type. In inhabited areas, more casualties are likely to result from a failed dam or a damaged nuclear reactor, for example, than from a damaged oil pipeline.

The factors which must be considered in assessment of seismic risk of a site have been well-defined in recent times<sup>(1-7)</sup>. Here a brief summary of these factors is listed.

*Geological Input* Any of the following investigations may be required.

- 1. Provision of a structural geologic map of the region, together with an account of recent tectonic movements.
- 2. Compilation of active faults in the region and the type of displacement (e.g., leftlateral, strike-slip, etc.). Fieldwork is sometimes necessary here. Of particular importance are geological criteria for fault movements in Holocene time (the past

10,000 years) such as displacements in recent soils, dating by radio-carbon methods of organic material in trenches across the fault, and other methods.

- 3. Mapping of the structural geology around the site, with attention to scarps in bedrock, effects of differential erosion and offsets in overlying sedimentary deposits. Such maps must show rock types, surface structures and local faults, and include assessments of the probable length, continuity and type of movement on such faults.
- 4. In the case of through-going faults near the site, geophysical exploration to define the location of recent fault ruptures and other lineaments. Geophysical work sometimes found useful includes measurement of electrical resistivity and gravity along a profile normal to the fault. Other key geological information is evidence for segmentation of the total fault length, such as step-over of fault strands, and changes in strike.
- 5. Reports of landslides, major settlements, ground warping or inundation from floods or tsunamis at the site.
- 6. Checks of ground water levels in the vicinity to determine if ground water barriers are present which may be associated with faults or affect the soil response to the earthquake shaking.

*Seismological Input* Procedures for the estimation of ground shaking parameters for optimum engineering design are still in the early stages and many are untested. It is important, therefore, to state the uncertainties and assumptions employed in the following methods:

 Detailed documentation of the earthquake history of the region around the site. Seismicity catalogs of historical events are needed in preparing lists of felt earthquakes. The lists should show the locations, magnitudes and maximum Modified Mercalli intensities for each earthquake. This information should be illustrated by means of regional maps.

- 2. Construction, where the record permits, of recurrence curves of the frequency of regional earthquakes, down to even small magnitudes (See the Gutenberg-Richter equation, Chapter 2). Estimates of the frequency of occurrence of damaging earthquakes can then be based on these statistics.
- 3. A review of available historic records of ground shaking, damage, and other intensity information near the site.
- 4. Estimation of the maximum Modified Mercalli intensities on firm ground near the site from felt reports from each earthquake of significance.
- 5. Definition of the design earthquakes<sup>(1-19)</sup>. The geological and seismological evidence assembled in the above sections should then be used to predict the earthquakes which would give the most severe ground shaking at the site. (Several such design earthquakes might be necessary and prudent.) Where possible, specific faults on which rupture might occur should be stated, together with the likely mechanism (strike-slip, thrust, and so on). Likely focal depth and length of rupture and estimated amount of fault displacement should be determined, with their uncertainties. These values are useful in estimating the possible magnitude of damaging earthquakes from standard curves that relate fault rupture to magnitude (see Table 1-6).

*Soils Engineering Input* When there is geological indication of the presence of structurally poor foundation material (such as in flood plains and filled tidelands), a field report on the surficial strata underlying the site is advisable. In addition, areas of subsidence and settlement (either natural or from groundwater withdrawal) and the stability of nearby slopes must be studied. We mention here only three factors that may require special scrutiny.

1. Study of engineering properties of foundation soils to the extent warranted for the type of building. Borings, trenchings and excavations are important for such analyses, as well as a search for the presence of sand layers which may lead to liquefaction.

- 2. Measurements (density, water content, shear strength, behavior under cyclic loading, attenuation values) of the physical properties of the soil in situ or by laboratory tests of borehole core samples.
- 3. Determination of P and S wave speeds and Q attenuation values and in the overburden layers by geophysical prospecting methods.

Table 1-6.	Earthquake	Magnitude	versus	Fault
	Ruptur	e Length		

	- 0	
Magnitude (Richter)	Rupture (km)	
5.5	5-10	
6.0	10-15	
6.5	15-30	
7.0	30-60	
7.5	60-100	
8.0	100-200	
8.5	200-400	

# 1.10 EARTHQUAKE AND GROUND-MOTION PREDICTION

Aspects of earthquake prediction that tend to receive the most publicity are: prediction of the place, prediction of the size, and prediction of the time of the earthquake. For most people, prediction of earthquakes means prediction of the time of occurrence. A more important aspect for mitigation of hazard is the prediction of the strong ground motion likely at a particular site (see also Chapters 2 and 3).

First, considering the status of forecasting of the time and size of an earthquake<sup>(1-20)</sup>. Prediction of the region where earthquakes are likely to occur has now been largely achieved by seismicity studies using earthquake observatories. Because empirical relations between the magnitude of an earthquake and the length of observed fault rupture have also been constructed (see Table 1-6), rough limits can be placed on the size of earthquakes for a region.

Many attempts have been made to find clues for forewarning. Some physical clues for earthquake prediction are shown in Figure 1-31. In 1975, Chinese officials, using in part, increased seismicity (foreshocks) and animal restlessness, evacuated a wide area before the damaging Haicheng earthquake. However, in the 1976 Tangshan catastrophe no forewarnings were issued. Elsewhere, emphasis has been placed on geodetic data, such as geodimeter measurements deformation of of the Californian crust along the San Andreas fault. An ex post facto premonitory change in ground level was found after the Niigata earthquake, which if it had been discovered beforehand, might have served as one indication of the coming earthquake.

Another scheme is based on detecting spatial and temporal gaps in the seismicity of a tectonic region. In 1973, a prediction was made by seismologists of the U.S. Geological Survey that an earthquake with a magnitude of 4.5 would occur along the San Andreas fault south of Hollister within the next six months. The prediction was based on four principal shocks which had occurred within 3 years on both ends of a 25-km-long stretch of the San Andreas fault, bracketing a 6-km-long section free from earthquakes in that time interval. The assumption was that the mid-section was still stressed but locked, ready to release the elastic energy in an earthquake. However, no earthquake occurred in the six months predicted. One difficulty with such methods is the assessment of a zero epoch with which to compare the average background occurrence rate.

A much publicized prediction experiment in California depended on the detection of a 22 year periodicity in moderate magnitude ( $M_L$  = 5.5) earthquakes centered on the San Andreas fault near Parkfield. Similar earthquakes were recorded in 1901, 1922, 1934, and 1966. Available seismograms in addition allowed quantitative comparison of the source mechanisms for the 1922, 1934, and 1966 earthquakes. Many monitoring instruments were put in place to try to detect precursors for a possible 1988 repetition. These included changes in ground water tables, radon concentration, seismicity, and fault slippage. The prediction of repetition of such a characteristic earthquake in the years  $1988 \pm 4$ , proved to be unsuccessful.



*Figure 1-31* Physical clues for earthquake prediction. [After predicting earthquakes, National Academy of Sciences. (1-1).]

It has often been pointed out that even if the ability to predict the time and size of an earthquake was achieved by seismologists, many problems remain on the hazard side. Suppose that an announcement were made that there was a chance of one in two of a destructive earthquake occurring within a month. What would be the public response? Would the major industrial and commercial work in the area cease for a time, thus dislocating large segments of the local economy? Even with shorter-term prediction, there are difficulties if work is postponed until the earthquake-warning period is over. Suppose that the predictive time came to an end and no earthquake occurred; who would take the

responsibility of reopening schools and resuming other activities?

Secondly, let us consider the calculation of artificial (synthetic) seismic strong ground motions $^{(1-21)}$ . The engineering demand is for the estimation of certain parameters, which will be used for design and structural checking. There are two representations usually used. The first is the seismogram or time history of the ground motion at the site represented instrumentally by the seismogram or accelerogram. The second is the Fourier or response spectra for the whole motion at the site. These two representations are equivalent and are connected by appropriate transformations between the time and frequency domains.

In the simplest time-history representation, the major interest is in the peak amplitudes of acceleration, velocity, and displacement as a function of frequency of the ground motion. Another parameter of great importance is the duration of the strong ground motion, usually given in terms of the interval of time above a certain acceleration threshold (say 0.05g), in a particular frequency range. Typically, the duration of a magnitude 7 earthquake at a distance of 10 kilometers is about 25 seconds. The pattern of wave motion is also important in earthquake engineering because the nonlinear response of structures is dependent on the sequence of arrival of the various types of waves, In other words, damage would be different if the ground motion were run backwards rather than in the actual sequence of arrival. In this respect, phasing of the ground motion becomes very important and the phase spectra should be considered along with the amplitude spectrum.

The phasing of the various wave types on synthetic seismograms can be determined by estimation of times of arrival of the P, S, and surface waves. In this way, a realistic envelope of amplitudes in the time histories can be achieved. There are two main methods for constructing synthetic ground motions. The first is the more empirical and involves maximum use of wave motion parameters from available strong ground motion records and application of general seismological theory<sup>(1-22)</sup>.

The second method entails considerable computer analysis based on models of the earthquake source and assumptions on earthquake scaling<sup>(1-16,1-23)</sup>. In the first method, the initial step is to define, from geological and seismological information, the appropriate



*Figure 1-32.* Ground motions recorded at Capitola, California in the 1989 Loma Prieta earthquake (a). The synthesized ground motions for site E23 (on rock) for a magnitude 7.2 earthquake, generated by shallow rupture 5 km away (b)



*Figure 1-33.* Response spectrum for the Capitola motion (Figure 1-32) compared with a target code spectrum (a). On the right, the ratio of frequency components is shown. Four iterations of the Fourier amplitude spectrum (phase fixed) of the Capitola record produced the compatibility shown (b).

earthquake sources for the site of interest. The source selection may be deterministic or probabilistic and may be decided on grounds of acceptable risk. Next, specification of the propagation path distance is made, as well as the P, S and surface wave velocities along the path. These speeds allow calculation of the appropriate wave propagation delays between the source and the multi-support points of the structure and the angles of approach of the incident seismic waves.

The construction of realistic motions then proceeds as a series of iterations, starting with the most appropriate observed strong motion record available, to a set of more specific time histories, which incorporate the seismologically defined wave patterns. The strong motion accelerograms are chosen to satisfy the seismic (dip-slip, etc.), and source type path specifications for the seismic zone in question. The frequency content is controlled by applying engineering constraints, such as a selected response amplitude spectrum<sup>(1-24,1-25)</sup>. The target spectra is obtained, for example, from previous data analysis, often from earthquake building codes. The fit between the final iteration and the target spectrum should fall within one standard error. Similarly, each

seismogram must maintain the specified peak accelerations. velocities ground and displacements within statistical bounds. The duration of each wave section (the P, S and surface wave portions) must satisfy prescribed source, path and site conditions. Figure 1-32 shows iterations of ground motion at site E23 for the east crossing of the San Francisco Bay Bridge for a Safety Evaluation Earthquake (SEE) of magnitude 7.2 on the nearby Hayward fault. The initial accelerogram chosen for the input motion at the pier in question is the horizontal ground motion recorded at Capitola on firm ground in the 1989 Loma Prieta earthquake. The north-south component of accelerations are shown. These records are then scaled for the required peak acceleration and then the 5% damped response spectrum is calculated. Two steps in the fitting of the response spectrum are shown in Figure 1-33. The ratio function between the calculated and target response spectrum is achieved. The process produces a realistic motion for the input site as shown in Figure 1-32. The synthetic motion has a wave pattern, including "fling", duration, amplitude and spectrum that are acceptable from both the seismological and engineering viewpoints.

At this stage, for large multi-support structures, account can be taken of the *incoherency* of ground motion. The first step is to lag, at each wave-length, the phase of the ground motion to allow for the different times of wave propagation between input points. Use must then be made of a coherency function (see Figure 1-34) that has been obtained from previous studies in similar geological regions. The process is to adjust the phase in the Fourier spectrum at each frequency so that the resulting phase spectrum for each input matches the selected coherency function.

Consider now the second method of synthesizing ground motions by computational means using a scaling relation between a small earthquake in the region and the ground motion required for engineering design or safety evaluation.

These are usually, of course, much smaller magnitude sources than required. Such smaller recorded ground motions contain essential particular earthquake properties of the mechanism involved, however, as well as the effects of the particular geological structure between the source and the station. In terms of the theory of the response of mechanical structures, they are called empirical Green's functions<sup>(1-16)</sup>. They can be considered as the response of the local geological system to an approximate impulse response of short duration applied at the rupturing fault. If such empirical Green's functions appropriate to the study of the site in question are not available, they must be constructed making certain mathematical assumptions and introducing appropriate tensor analysis (see Section 1.8).



*Figure 1-34.* The coherency function commuted from adjacent (1750m separation) recordings of strong ground motion in the Gilroy strong ground motion array during the 1989 Loma Prieta main shock. The heavy line is a smooth representation of the coherency effect.

The size of the earthquake is then selected in terms of its seismic moment which, as was seen in Section 1.8, is given quantitatively in terms of the area of the fault slip and the amount of slip. An appropriate fault area is then mapped in terms of an elementary mesh of finite elements. The empirical Green's function

#### Acceleration Time Histories



*Figure 1-35.* Acceleration ground motion recorded at Sylmar, California in the 1994 Northridge earthquake (a). Acceleration ground motion recorded at Pacoima Dam in the 1971 San Fernando earthquake (b).

mentioned above is then applied to each element of the mesh in the sequence required to achieve the appropriate rupture velocity across the whole fault surface, as well as maintain the specified overall moment for the larger earthquake. This superposition can be done in terms of amplitude and phase spectra by available computer programs and the synthetic seismogram calculated at a point on the surface in the vicinity of the original recorded Green's function. empirical Various modifications of the process described above have been explored and a number of test cases have been published.

As can be seen, the prediction of ground motions in this way involves a number of assumptions and extrapolations. A particular value of the method is to permit the exploration of the effect of changing some of the basic parameters on the expected ground motions. A major difficulty is often the lack of knowledge of the appropriate wave attenuation for the region in question (see Section 1.5). Because of the importance of the application of attenuation factors in calculation of predicted ground motion at arbitrary distances, a great deal of work has been recently done on empirical attenuation forms<sup>(1-26,1-27)</sup>. The usual form for the peak value at distance x is given by:

$$y = \frac{ae^{DM}}{ce^{dM} + x^2}$$

Where a, b, c, and d are constants and M is the magnitude. Recent empirical fits are given in Reference 1-31.

As an example of the type of attenuation curve that is obtained from actual ground

motion recordings, a relation for the peak ground displacements, D, in the 1989 Loma Prieta and 1992 Landers earthquake is given in the following equation.

 $\log D = 1.27 + 0.16 \text{ M} - \log r + 0.0004 \text{ r}$ 

Where D (cm) is the displacement, M is the moment magnitude, and r (km) is the distance to the nearest point of energy release on the fault.

It is usual that attenuation changes significantly from one geological province to another and local regional studies need to be made to obtain the parameters involved. A discussion is given in the book by Bullen and Bolt<sup>(1-12)</sup> and in Reference 1-28.

The Northridge earthquake of California, January 17, 1994 allows an important comparison between theoretical the seismological expectations and actual seismic wave recordings and behavior of earthquake resistant structures. This magnitude 6.8 earthquake struck southern California at 4:31 AM local time on January 17, 1994. The earthquake rebound occurred on a southerly dipping blind-thrust fault (see Section 1.4). The rupture began at a focus about 18 km deep under the Northridge area of the San Fernando Valley. The rupture then propagated along a 45° dipping fault to within about 4 km of the surface under the Santa Susannah Mountains. No major surface fault rupture was observed although the mountainous area sustained extensional surface fracturing at various places and were uplifted by tens of centimeters. The causative fault dipped in the opposite sense to that which caused the neighboring 1971 San Fernando earthquake.

Like the 1971 earthquake, the 1994 shaking tested many types of design such as base isolation and the value of the latest Uniform Building Codes. Notable was, again, the failure of freeway bridges designed before 1971 and the satisfactory seismic resistance of Post-1989 (Loma Prieta earthquake) retrofitted freeway overpasses. The peak accelerations, recorded by many strong motion accelerometers in Los Angeles and the San Fernando Valley area, were systematically larger than average for average curves obtained from previous California earthquakes. It is notable that the ground motions at the Olive View Hospital (see Figure 1-35) are similar to those obtained at the Pacoima Dam abutment site in the 1971 thrust earthquake <sup>(1-29)</sup>.

## 1.11 CONCLUSIONS

The state of the art in strong motion seismology is now such that prediction of key parameters, such as peak ground acceleration and duration of the significant portion of shaking at a given site, is relatively reliable. Recent earthquakes, such as Chi Chi, Taiwan 1999, have provided many recordings of the strong motion and various site conditions of rock and soil. In addition the great earthquakes of 1985 in Chile and Mexico ( $M_s \approx 8$ ) yielded accelerograms for large- subduction-zone earthquakes. There are still, however, no clear recordings of ground motion in the near-field from earthquakes with  $M_s > 7.5$  so that extrapolations to synthetic ground motions in extreme cases of wide engineering interest are not available.

To meet this need and others of engineering importance, more strong motion instruments are being placed in highly seismic areas of the world. Of special interest is the recent operation of clusters of digital instruments in urban areas. These allow the intensity of seismic waves involved in strong motion shaking in the near-field to be rapidly computed and distributed on the world wide web (see <u>http://www.trinet.org</u>).

Of special importance is the instrumentation of large structures (such as large dams and long bridges). But structural analysis requires realistic predictions of free field surface motions at all interface points on the supporting foundation under design earthquake conditions. In the past, engineers have normally carried out seismic analyses under the incorrect assumption that the motions of all support points are fully correlated, i.e. "rigid foundation" inputs are assumed<sup>(1-30)</sup>.

As recent strong motion data have come to hand, these observations allow the study of the effects of magnitude, epicentral distance, focal depth, etc. on such characteristics<sup>(1-31)</sup>.

The seismological problems dealt with in this chapter will no doubt be much extended in subsequent years. First, greater sampling of strong-ground motions at all distances from fault sources of various mechanisms and magnitudes will inevitably become available. An excellent example is the wide recording of the 1999 Chi-Chi, Taiwan, earthquake<sup>(1-32)</sup>. Secondly, more realistic three dimensional numerical models will solve the problem of the sequential development of the wave mixtures as the waves pass through different geological structures. Two difficulties may persist: the lack of knowledge of the roughness distribution along the dislocated fault and, in many places, quantitative knowledge of the soil, alluvium, and crustal rock variations in the region.

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