NATURE OF EARTHQUAKES

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Summary

Earthquakes are one of the worst natural calamities. According to conservative estimates, in recorded history, tens of millions of people have lost their lives in earthquakes and damage has run into hundreds of billions of dollars. In the recent past, the 1976 Tangshan earthquake in the People's Republic of China claimed 242 000 human lives and is qualified to be the deadliest twentieth-century earthquake. The 1995, the Kobe earthquake in Japan was the most expensive natural disaster and is estimated to have caused economic losses of US$150–200 billion. The first section in the paper is introductory and provides the necessary definitions and information about earthquakes. The second section provides with detailed information about some of the important earthquakes of the twentieth century. With the passage of time, it is being realized that earthquakes can be triggered by of anthropogenic activities. Among the triggered earthquakes, the ones triggered by filling of artificial water reservoirs are the most significant. The third section deals with these. In the fourth section, the elements of seismic zoning are discussed and a summary information is given on the recently concluded Global Seismic Hazard Assessment Program which has provided unified basic information on the probable seismic hazard for the entire world. Earthquake prediction has been a topic of interest for centuries. Even at the beginning of the twenty-first century, we do not have a short-term earthquake prediction available. However, a lot of progress has been made and there are indications that limited prediction may be possible in the next two or three decades. The last addresses these issues.

1. Basics of Seismology

Simply stated, shaking or convulsion of Earth is called an earthquake. Earthquakes result from the buildup of stresses within the rocks until the accumulated strain exceeds their strength and fracture occurs. They radiate seismic waves of various types, which propagate in all directions through earth's interior. The passage of seismic waves through rocks cause shaking that we feel as earthquakes. Majority of earthquakes are tectonic earthquakes associated with faults in rocks. Even among these, most are due to reactivation of preexisting faults and the rest due to fresh faulting. Some transient ground disturbances in a volcanic region are due to explosive release of gases and others to movement of magma in volcanic pipes. Volcanic tremors are rapid successions of transient disturbances in volcanic regions. Landslides, the collapse of caves, and meteorite impacts are natural but rare cases of transient ground disturbances. Nuclear and chemical explosions, mining, fluid injection, and reservoir-triggered earthquakes are humanmade transient disturbances.

To understand and appreciate earthquakes, it is important to learn about the structure of Earth and the concept of plate tectonics. In the following subsections, we first define the most commonly used terms in studying earthquakes along with a few other definitions and then briefly describe the structure of Earth, plate tectonics, the magnitude and intensity of earthquakes, and so forth.

1.1. Earth's Structure

Earth's radius is 6371 km and its structure can be approximated by a series of concentric spherical shells as shown in Figure 1. This figure shows the large-scale features of Earth's internal structure. The two innermost regions constitute the core, which has the greatest average density, exceeding 10 g cm^{-3}, with iron-nickel alloy being the most likely constituent. The outer part of the core is molten and the inner core is solid. This is inferred from the transmission of seismic waves, as the outer part of the core does not transmit shear waves. The core is covered by the mantle with an average density of 4.5 g cm^{-3}, indicating that it is constituted of rocks rather than metals. Based on its density, seismic wave velocity, and the study of rocks that are believed to have come from the mantle, the mantle is inferred to be predominantly constituted of oxygen and silicon with magnesium and iron as the most abundant metallic ions. On the basis of the propagation of earthquake waves, the mantle between the core and crust is divided into the lower mantle, transition zone, and upper mantle (Figure1). As a result of increase in pressure, the seismic velocity and density increase with depth in the lower mantle. The increase in the amount of iron in the silicate minerals with depth is responsible for increase in density and velocity. The upper mantle is mainly composed of olivine, pyroxene, and garnet. The Mohorovicic discontinuity, named after its discoverer in the early twentieth century (also known as the Moho) separates the crust from the upper mantle. The composition of the crust beneath the ocean differs from that of continents. The oceanic crust accounts for about 65% of Earth's surface and it is covered with water with an average depth of 4 km. Below water, on average, there are about 0.5 km of sediments overlying an approximately 1.5 km thick layer of basaltic volcanic rocks. The next 6 km down to Moho has primarily metamorphosed basaltic volcanic rocks and other iron- and magnesium-rich igneous rocks. In contrast to a relatively uniform oceanic crust, the continental crust varies from a thickness of less than 25 km under certain shield areas to more than 70 km below certain high mountains like the Himalayas and the Tibet Plateau, with an average thickness of 35 km. The crust has a very complicated structure. At the surface, a great variety of rocks are exposed, including sediments, shale, sandstones and limestones, and the ancient shields are mostly composed of granite and volcanic lava. The underlying basement is comprised of granites. In addition to basic differences between continental and oceanic crusts, such as elevation, thickness, structure, and overall composition, the transition from ocean to continent provides very interesting situations.

Figure 1. The structure of Earth with an average radius of 6371 km.
Cr: crust, thickness 5 to 70 km; Mn: mantle extending to a depth of 2900 km comprising

upper mantle (m_1), transition zone (m_2), and lower mantle (m_3), and Co: the core comprising of an outer liquid part (C_1) and the inner solid core (C_2).

1.2. Plate Tectonics

The concept of plate tectonics was developed during the 1960s. Seismology has provided much evidence in support of this concept. In turn, almost all our observations and perceptions about earthquakes so far can be rationalized within the plate-tectonic hypothesis. Tectonics is the study of deformation of Earth's material and when we discuss it on a global scale, we call it plate tectonics. There are some very simple basic assumptions. The first assumption is that the outer portion of Earth acts as a rigid cap or plate on a sphere. These plates consist of the crust and a part of the upper mantle and are about 100 km thick, also known as the lithosphere. There are seven major and several small plates on Earth's surface (Figure 2). These plates do not undergo any significant internal deformation. The second assumption is that each plate is in relative motion with respect to other plates on the surface of Earth and that any significant deformation occurs only at the plate boundary. This deformation causes stress and when the accumulated strain exceeds the strength of the rocks, earthquakes occur. To explain this in a very simple way, if we take a piece of chalk a few inches long and try to bend it, we notice that it does not bend because the chalk is hard. As a matter of fact, the chalk is a sort of rock. But if we press the ends harder, the chalk snaps suddenly into two pieces. Something similar occurs during earthquakes. A majority of earthquakes occur at the plate boundaries. The plate boundaries are divided into convergent, divergent, and transform plate boundaries. The divergent plate boundaries, such as mid-oceanic ridges, are the places where the new crust is created continuously whereas at convergent plate boundaries, such as subduction zones, it is consumed. At transform plate boundaries, two plates slide past each other.

Figure 2. Major plates, identified by their names, on Earth's surface
Dots indicate earthquake epicenters of magnitude >5 that occurred during 1973–1999. The three major seismic belts (P, Circum-Pacific belt; A, Alpide Himalayan belt; and O, mid-oceanic ridges) are also identified in this figure. Note the correlation between the plate boundaries and seismicity.

1.3. Earthquake Belts

During the 1960s, with the deployment of the World-Wide Standard Seismograph Network (WWSSN), where about 120 seismic stations with similar equipment were deployed globally, earthquake location capabilities improved by severalfold and the plate boundaries, in the form of narrow belts of earthquakes, were recognized (Figure 2). Basically, there are three major belts that account for almost 95% of earthquake activity. The belt along which most of the earthquakes occur is called the Circum-Pacific belt which goes around the rim of the Pacific Ocean right from the southern tip of South America through Central America, California, the Aleutian Islands, Japan, and down to New Zealand. The second most active belt is the Alpide-Himalaya seismic belt which starts from southeast Asia near Java-Sumatra, continues through Andaman

Nicobar Islands, India-Burma border region, swings through north of India in the foothills of Himalaya and then moves west through Iran into Greece and Italy. The third major seismic belt consists of mid-oceanic ridges, which account for small-magnitude earthquakes. These can be seen in the Pacific, Atlantic, and Indian Oceans. Globally, hundreds of earthquakes occur annually. Approximate annual frequency of earthquake occurrence in the world is given in Table 1.

Table 1. Annual frequency of earthquakes by magnitude

A majority of these earthquakes occur at shallow depths (0–70 km), some occur at intermediate depths (70–300 km), and a few at deeper depths (300–700 km).

1.4. Faults

A fault is a fracture or a fracture zone along which there has been displacement of the two sides relative to one another parallel to the fracture. As noted earlier, earthquakes occur on fresh or preexisting faults. The sense of motion on the fault in an earthquake is governed by the orientation of the fault and the plate tectonic forces. Geologists identify three important categories of faults (Figure 3). A strike-slip fault is vertical and the rocks across it move horizontally relative to each other. A normal fault is inclined and the rocks above it move down relative to those below it. A reverse fault is also inclined but the rocks above it move up relative to those below it. Subhorizontal reverse faults are common and are called thrust faults. Faults in which normal and strike-slip motions occurr together or reverse/thrust and strike-slip motions occurr together have also been observed. These faults exist in all parts of the lithosphere. Earthquake-generating faults are most abundant along the plate boundaries. Some active faults occur in plate interiors also. Analyses in practical cases show that earthquakes may be associated with faults of all the above types. But earthquakes with reverse/thrust-type fault motion at the hypocenters are common at those plate boundaries where adjacent lithospheric plates move towards each other (e.g., in the Himalaya). Earthquakes with normal-type fault motion at the hypocenters are common at divergent plate boundaries (e.g., along mid-oceanic ridges). Earthquakes with strike-slip-type fault motion at the hypocenters are common where the lithospheric plates slip past each other horizontally (e.g., along the San Andreas fault).

Figure 3. Three major types of faulting
FF indicates the fault plane. In normal faulting, the block above the fault moves down whereas in thrust faulting the block above the fault moves up. Strike-slip faulting involves sliding past of the two blocks along the fault plane FF without any vertical movement.

1.5. Seismic Waves

If we are close to the epicenter of an earthquake, we feel a sudden jolt that is followed by shaking of the ground after a few seconds depending upon the distance of the earthquake's hypocenter. What we experience are two kinds of waves. The jolt is caused by the body waves, which travel through the ground and reach us, whereas the shaking is caused by the surface waves, which travel along the surface of the ground. The surface waves can also be imagined by comparing these with dropping of a stone in calm pool of water which generates the ripples spreading from the point where the stone has fallen with the passage of time. There are two main types of surface waves. Rayleigh waves have retrograde elliptical particle motion in the direction of propagation, whereas in Love waves, the particle motion is at a right angle to the direction of propagation. The body waves are generated by fracturing of rocks and there are basically two types of body waves: primary (P), known as longitudinal waves, and secondary (S) or shear waves. In the case of P waves, the particles vibrate to and fro along the path of wave propagation from the source. They are comparable to sound waves in air. The S waves represent transverse elastic waves in which particles vibrate perpendicular to the path. These are comparable to transverse waves on a string. S waves propagate only in solids.

1.6. How We Record and Locate Earthquakes

We record earthquakes with instruments called seismographs. Broadly speaking, a seismograph is an instrument that continuously monitors and, in some form, records ground vibrations continuously as a function of time. Ground vibrations are caused by the arrival of seismic waves. It also keeps a very precise information on the timing so that we know when exactly a wave has arrived. The main component of a seismograph is a seismometer, which makes it possible to record the mechanical energy of the arriving waves. In the earliest developed seismometers, this was done mechanically by enhancing the movements. The simplest form of seismometers were developed by the end of the nineteenth century by Milne, Grey, and Ewing at the Imperial College of Engineering in Tokyo. Figure 4 shows the principle. Basically, a seismometer employs a pendulum. The pendulum mass is connected through a vertical or horizontal suspension to a frame, which in turn is anchored to the ground. When the seismic wave reaches the recording site it causes the frame to move along with the ground, whereas the mass, due to its inertia and loose coupling to the frame, tends to remain stationary. The relative motion between the mass and the frame can be magnified through various methods. In earlier models, the movement was recorded optically on a strip of photographic paper. Later, transducers were used, which converted the mechanical energy into electric voltage. A typical record of an earthquake showing the primary, secondary, and surface waves is shown in Figure 5.

Figure 4. Principle of the seismometer
The relative motion between the recording drum and the frame of instrument to which the pen is attached helps in recording the earthquake.

Figure 5. A seismogram recorded at the National Geophysical Research Institute seismic station, Hyderabad, India, for an earthquake of magnitude 5.4 located at a distance of about 2700 km in the Chagos Archipelago region
The beginning of the P (primary or longitudinal), S (secondary or transverse), and SW (surface) waves is marked. The arrival of P and S waves is separated by about 5 minutes, corresponding to a distance of 2700 km.

The P and S waves travel with different velocities in the earth, the P-wave velocity being a little less than two times the S-wave velocity. Since both these waves are generated at the focus at same time, as they travel longer distances as the time interval between the arrival of P and S waves increases. Therefore, at a given seismic station, by measuring time of arrival of P and S waves, we can determine the distance these waves have traveled before being recorded at that station. However, this earthquake could have occurred at that distance in any direction from the station. It must be mentioned that all seismic stations in the world run on the same universal time. If this earthquake is recorded at least two other stations, we may draw three circles corresponding to their difference in P and S time intervals. The place where the three circles intersect is the epicenter. This is a very simple way of demonstrating how an earthquake is located. Globally, hundreds of stations record the same earthquake and the information is exchanged to accurately determine the focal parameters of the earthquakes.

1.7. Earthquake Magnitude and Intensity

The magnitude of an earthquake is a measure of the amount of energy released at the hypocenter, whereas the intensity tells the effect of this earthquake at a given location. Although several seismologists have made efforts to define earthquake magnitude, the real quantitative and globally acceptable earthquake magnitude scale was defined by Charles Richter in 1935. Professor Richter borrowed the term magnitude from astronomy where the brightness of a star is measured on a logarithmic scale and referred to as its magnitude. As defined by Richter in his own words "The magnitude of any shock is taken as the log of the maximum of the trace amplitude, expressed in microns, with which the standard short period torsion seismometer would register that shock at an epicentral distance of 100 km." This scale was basically developed to measure the magnitudes of earthquakes in Californian region. At that time the Wood-Anderson torsion seismometers were in vogue. It may also be mentioned that certain calibration is required as a record on a Wood-Anderson seismometer may not always be available at a distance of 100 km from the epicenter. What is important is to know the distance of an earthquake from the recording station and also to measure the amplitude of the maximum P-wave motion on the seismogram. By correlating the motion on the seismogram and the distance, the earthquake magnitude is estimated. With the passage of time, many more magnitude scales have been developed since the original Richter scale was limited to a small area and to one particular instrument. For example, $M = \log_{10}(A/KT) + \Delta + S$. In this equation, M is the magnitude, A is amplitude recorded on the seismogram in mm, K is the magnification of the instrument in thousands, T is the period of the wave in seconds, Δ is the depth distance factor in degrees, and S is the station factor. Today the most commonly used magnitude scales are the body wave magnitudes, which make use of the P-waves (m_b). M is the surface wave magnitude used worldwide and is derived by using the Rayleigh waves of the 20-second period.

Energy released in an earthquake increases with its magnitude. It is estimated that the total energy released by 32 earthquakes of magnitude M, say, equals that released by one earthquake of magnitude M + 1. The largest earthquakes have energy equivalent to about 12 000 atomic bombs of the type dropped on Hiroshima.

The procedure for describing the effect of an earthquake in a locality has been systematized through the development of earthquake intensity scales. Intensity is a qualitative assessment of the damage experienced at a point due to a given earthquake. Intensity is assigned by human observers and is thus a subjective measure. The Modified Mercalli (MM) intensity scale is in use almost universally today. In it, the permanent and transient macroscopic effects of earthquakes are graded and divided into ordinal classes I through XII. Intensity MM XII is assigned where the most violent effects of earthquakes are observed. The Modified Mercalli intensity scale is given below.

I. Not felt. Marginal and long period effects of large earthquake.
II. Felt by persons at rest, on upper floors, or favorably placed.
III. Felt indoors. Hanging objects swing. Vibration like passing of light trucks. Duration estimated. May not be recognized as an earthquake.
IV. Hanging objects swing. Vibration like passing of heavy trucks; or sensation of a jolt like a heavy ball striking the walls. Standing motor cars rock. Windows, dishes, doors rattle. Glasses clink. Crockery clashes. In the upper range, wooden walls and frame creak.
V. Felt outdoors; direction estimated. Sleepers wakened. Liquids disturbed, some spilled. Small unstable objects displaced or upset. Doors swing, close, open. Shutters, pictures move. Pendulum clocks stop, start, and change rate.
VI. Felt by all. Many frightened and run outdoors. Persons walk unsteadily. Windows, dishes, glassware broken. Knickknacks, books, etc., fall off shelves. Pictures off walls. Furniture moved or overturned. Weak plaster and masonry D cracked. Small bells ring (church, school). Trees, bushes shaken (visibly, or heard to rustle).
VII. Difficult to stand. Noticed by drivers of motor cars. Hanging objects quiver. Furniture broken. Damage to masonry D, including cracks. Weak chimneys broken at roof line. Fall of plaster, loose bricks, stones, tiles, cornices (also unbraced parapets and architectural ornaments). Some cracks in masonry C. Waves on ponds; water turbid with mud. Small slides and caving in along sand or gravel banks. Large bells ring. Concrete irrigation ditches damaged.
VIII. Steering of motor cars affected. Damage to masonry C; partial collapse. Some damage to masonry B; none to masonry A. Fall of stucco and some masonry walls. Twisting, fall of chimneys, factory stacks, monuments, towers, and elevated tanks. Frame houses moved on foundations if not bolted down; loose panel walls thrown out. Decayed piling broken off. Branches broken from trees. Changes in flow or temperature of springs and wells. Cracks in wet ground and on steep slopes.
IX. General panic. Masonry D destroyed; masonry C heavily damaged, sometimes with complete collapse; masonry B seriously damaged. General damage to foundations. Frame structures, if not bolted, shifted off foundations. Frames racked. Serious damage to reservoirs. Underground pipes broken. Conspicuous cracks in ground. In alluvial areas, sand and mud ejected; earthquake fountains and sand craters occur.
X. Most masonry and frame structures destroyed with their foundations. Some well-built wooden structures and bridges destroyed. Serious damage to dams, dikes, embankments. Large landslides. Water thrown on banks of canals, rivers, lakes, etc. Sand and mud shifted horizontally on beaches and flat land. Rails bent slightly.
XI. Rails bent greatly. Underground pipelines completely out of service.

XII. Damage nearly total. Large rock masses displaced. Lines of sight and level distorted. Objects thrown into the air.

Definitions of terms used in this description are:
Masonry A: Good workmanship, mortar, and design; reinforced, especially laterally, and bound together by using steel, concrete, etc.; designed to resist lateral forces.
Masonry B: Good workmanship and mortar; reinforced, but not designed in detail to resist lateral forces.
Masonry C: Ordinary workmanship and mortar; no extreme weaknesses like failing to tie in at corners, but neither reinforced nor designed against horizontal forces.
Masonry D: Weak materials, such as adobe; poor mortar; low standards of workmanship; weak horizontally.

2. Significant Earthquakes

Several compilations exist on the significant earthquakes of the world. We have included in Table 2 a list of significant earthquakes, their magnitude and/or intensity, and the number of human lives lost. Until the beginning of the 21st century, the 1556 China earthquake is the deadliest with an estimate of 830 000 lives lost. In the twentieth century, the Tangshan, China, earthquake of 1976 is the deadliest having claimed 242 000 human lives and the Kobe, Japan, earthquake of 1995 is the most expensive one with the financial losses estimated to be between US$150–200 billion. A few of the important earthquakes of the twentieth century are discussed in chronological order in the following subsections.

Table 2. Notable Earthquakes of the World

2.1. 1906—San Francisco Earthquake, USA

The April 18, 1906, San Francisco earthquake on San Andreas fault is one of the most significant earthquakes of all time. At around 0512 local time, a foreshock occurred which was felt widely throughout the San Francisco area and this was followed some 20 to 25 seconds later by the main shock with the epicenter near San Francisco. The earthquake caused very violent shaking, which was estimated to last for about one minute, and the earthquake was felt at long distances. The maximum intensity on Modified Mercalli scale varied from VII to IX parallel to the length of the entire rupture, and extending up to 80 km inland from the fault trace. It was noted that the structures situated on sediment-filled valleys suffered much stronger shaking than those situated on nearby bedrock sites, while the maximum shaking was felt in San Francisco Bay area on the reclaimed grounds.

The original estimate of 700 deaths due to this earthquake is now questioned and it is estimated that the lives lost may be higher by a factor of three or four. Most of the casualties occurred in San Francisco area. The financial loss is estimated to be US$400 million (1906 dollars). It is also estimated that if such an earthquake was to occur today, several thousand people would have been killed and the economic losses would have run into hundreds of billions of dollars. The earthquake caused a great fire. The fire could not be contained for three days as the displacement on the San Andreas fault put the water pipelines out of operation. The fire was finally stopped by blasting structures in the path of the spreading fire.

Originally, a magnitude of 8.3 was assigned to this earthquake. However, recent work has down-sized the earthquake magnitude to 7.8. The rupture of the 1906 earthquake is estimated to be about 470 km in length. In comparison, the rupture of the 1989 Loma Prieta earthquake was only 45 km (Figure 6). The maximum displacement was about 7.5 m near Shelter Cove and Point Reyes (Figure 7).

Figure 6. The diagram shows 470 km long rupture of the 1906 San Francisco earthquake in the State of California.
For comparison, the rupture caused by the recent Loma Prieta earthquake of 1989 is also shown, which is only 40 km.

Figure 7. The diagram depicts the strike-slip movement on the San Andreas Fault at various locations.
The slippage due to the great 1906 San Francisco earthquake on the San Andreas Fault was greatest at Shelter Cove.

One of the significant outcomes of several investigations that were carried out after the San Francisco earthquake was the development of elastic rebound theory by Dr. Henry Fielding Reid, Professor of Geology at John Hopkins University, Baltimore, Maryland, USA which now forms the basic tenet of earthquake-occurrence models. The principle of the elastic rebound theory is illustrated in Figure 8. A straight fence built across the San Andreas fault would stretch as the Pacific Plate moves northwest. Just before the earthquake the fence will have an “S” shape. When the earthquake occurs the distortion is released and the two parts of the fence are straight again. However, there is an offset. Another very interesting observation made is that the 1906 earthquake essentially turned off earthquakes of magnitude 6 and larger for the next 73 years, with one exception of a 1911 earthquake, whereas there had been at least 16 earthquakes of magnitude 6 and larger in a period of 70 years before the 1906 earthquake (Figure 9). The reason for this has been ascribed to decrease in stress on the nearby faults due to the occurrence of the earthquake.

Figure 8. The elastic rebound theory was developed by H.F. Reid after the observations made during the San Francisco earthquake.
(a) A fence across the San Andreas Fault before the accumulation of strain; (b) the fence bends and takes S shape due to accumulation of strain; (c) during the earthquake the rupture causes movement and the two portions of the fence are displaced.

Figure 9. The diagram shows the spatiotemporal distribution of earthquakes during the period 1836 to 2001.
It is inferred that the 1906 San Francisco earthquake turned off earthquakes of magnitude 6 and larger along the San Andreas Fault.

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Biographical Sketch

Harsh Gupta is a well-known seismologist. His work on reservoir triggered earthquakes and study of physics of the earthquakes is globally appreciated. He has authored three books, all published by Elsevier Scientific Publishing Company in Amsterdam, and written over 130 papers published in International Journals. He has occupied several important positions globally. Currently, he is Secretary to the Government of India for the Department of Ocean Development. He was the Director of the National Geophysical Research Institute, Hyderabad, India and Adjunct Professor, University of Texas at Dallas, USA. He was the Founder President of the Asian Seismological Commission (ASC); Vice-President of the International Association of Seismology and Physics of Earth's Interior (IASPEI), and the Chairman of the Steering Committee of the Global Seismic Hazard Assessment Program (GSHAP). He is currently a Bureau Member of the International Union of Geology and Geophysics (IUGG) and Councilor of the International Union of Geological Sciences (IUGS).