Applied Seismology for Engineers

Dr. Abhishek Kumar

Department of Civil Engineering

Indian Institute of Technology Guwahati

Week – 04 Lecture - 02

Lecture – 08

Hello everyone, welcome to lecture 8 of the course Applied Seismology for Engineers, myself Dr. Abhishek Kumar. In lecture 7, we discussed about different kinds of seismic waves, which are generated at the source, primarily the P waves as well as the shear waves. When these waves interact with surficial and near surface mediums, again waves, primarily Rayleigh wave and Love waves, come into the picture. We also discuss the characteristics of the medium through which these waves can propagate, and also, we discussed in lecture 7 that when these waves are passing through a particular medium, what kind of particle motion is generated.

Recalling that when primary waves are passing through a particular medium, there will be compression and rarefaction. There will be particle motion happening in the longitudinal direction. When a shear wave is passing through a particular medium, it will cause back-and-forth motion in the perpendicular direction or shearing in the perpendicular direction, as a result of which there will be shearing happening in the perpendicular direction in the horizontal plane as well as in the vertical plane. Collectively, if you are interested in finding out the shearing happening in a particular plane, so that can be approximated by means of the application of torque on a particular section.

We also discussed primary wave and secondary wave shear zones, shadow zones, which are primarily because of significant change in the physical properties of the medium, and depending upon whether the medium is offering resistance to the propagation of wave or not. Some waves will be able to propagate through solid liquid also, other waves will not be able to propagate through liquid as well as gas. As a result, there will be shadow zones where, with respect to the epicenter, there will be some azimuth range within which you will not have any primary wave. Similarly, there will be some range of azimuth with respect to the epicenter where there will not be any shear waves. So, as a result of this, there will be shadow zones for primary waves as well as shear waves.

In addition, we discussed ground rolls, which are particularly the characteristics when a Rayleigh wave is passing through a particular medium, representing elliptical motion in the particle. We also discussed that the amplitude of Rayleigh wave as well as Love wave significantly reduces as we move from the ground surface to deep depths. If you go to Love waves, there will be motion completely in the horizontal plane. There will not be any motion in the vertical plane. Then, as we discussed, these ground motion recordings or seismic waves will be reaching different recording stations.

So, it will be starting from the source and will start propagating in all the directions. If we are having a recording station, the wave will reach the recording station, and thus the recording station will sense the characteristics of the wave as these change with respect to time. If you start analyzing the seismogram, that is, the ground motion signature recorded by a recording station, one is also able to understand where the primary wave content is coming into the picture, where the shear wave is coming into the picture, and accordingly, we can utilize this information, that is, the arrival time of the primary wave, the arrival time of the shear wave, which is marked at a recording station by means of a ground motion record. We use it to understand and locate primarily the location of the earthquake, as we generally target to locate the epicenter of the earthquake. We also discussed in yesterday's lecture 7 that we have to have such records of the arrival time of primary wave and secondary wave from at least three recording stations.

If it is less than three recording stations, then we may narrow it down to some area which is significantly larger than the actual area of epicenter location, which can be narrowed down if a record from more recording stations is there. So, referring to that question, the methodology which we discussed in lecture 7, we will be solving by numerical first, and then we will proceed towards other topics which are primarily related to earthquake intensity as well as magnitude, and in the end, we will also touch upon what is the seismic wave attenuation happening during the propagation path primarily. So, taking into account the formula which we discussed, if we know the time of primary wave or secondary wave reaching a recording station and velocity, that can be correlated with respect to the distance from which the primary wave or secondary wave was generated. But in this particular case, since the actual time of arrival is also dependent upon when the earthquake wave has started from the source and with respect to that, how much delay when the wave is reaching a particular recording station. So, many times, getting exact information about the arrival time of primary wave as well as secondary wave is difficult.

- **S-P formulae**

$$\mathbf{Time} = \frac{\mathbf{Distance(D)}}{\mathbf{Velocity}}$$

Time for Shear Wave or S-wave (V_s) Recorded (T_s)

$$\mathbf{T_s} = \frac{\mathbf{D}}{\mathbf{V_s}} \tag{1}$$

Time for Primary Wave or P-wave (V_p) Recorded (T_p)

$$\mathbf{T_p} = \frac{\mathbf{D}}{\mathbf{V_p}} \tag{2}$$

So, if we continue this particular derivation, similarly, we can also derive the time for shear wave arrival at a recording station, that is, T_s equals to D over V_s, where T_s is the arrival time of shear wave, D is the distance. In this particular case, this distance represents the epicentral distance, and V_s is the shear wave propagation velocity from a particular medium. Here, whenever we are talking about medium, primarily we are focusing on crystal medium through which between the source, that is the focus, and under the recording station again, at maybe weathered rock or maybe intact rock medium, the maximum content of the wave will be transferred, and then subsequently, these will be interacting with near-surface material. So, we will be targeting with respect to the crystal velocity of primary wave and secondary wave, and then the time for primary wave similarly can be obtained using this as T_p equals to D over V_p, where T_p is the time of arrival of primary wave, D is the epicentral distance, and V_p is the wave propagation velocity of primary wave in the medium of interest. So, using these as we discussed in lecture 7, if we take the difference between the two arrival times, definitely the arrival time of shear wave will be more in comparison to primary wave because, in comparison to primary wave, shear waves travel at relatively low value.

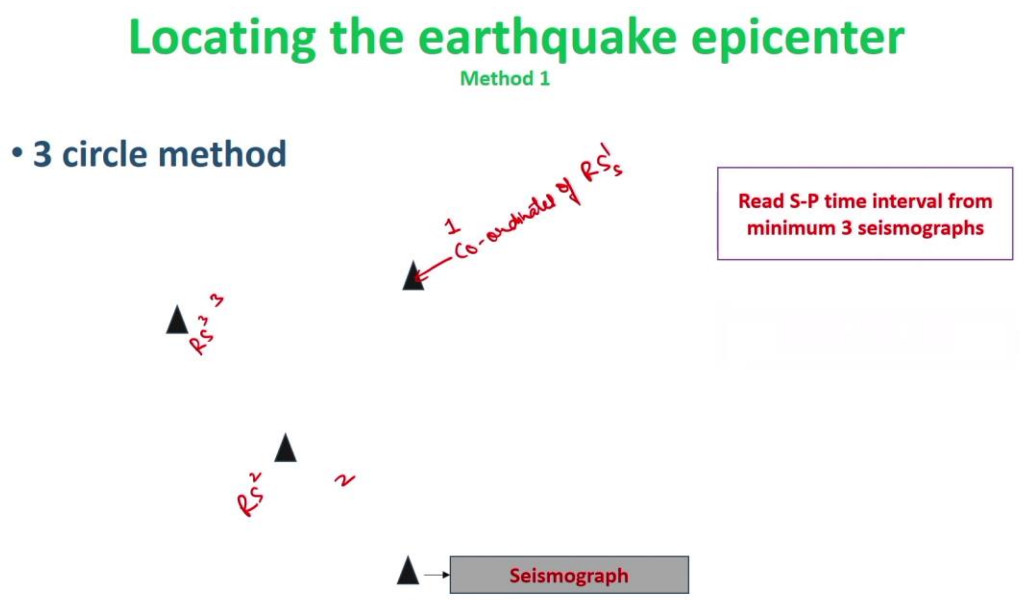
- **Subtracting equation 2 from equation 1**

$$\mathbf{T_s - T_p = D \times \frac{(V_p - V_s)}{(V_p \times V_s)}} \qquad \mathbf{3}$$

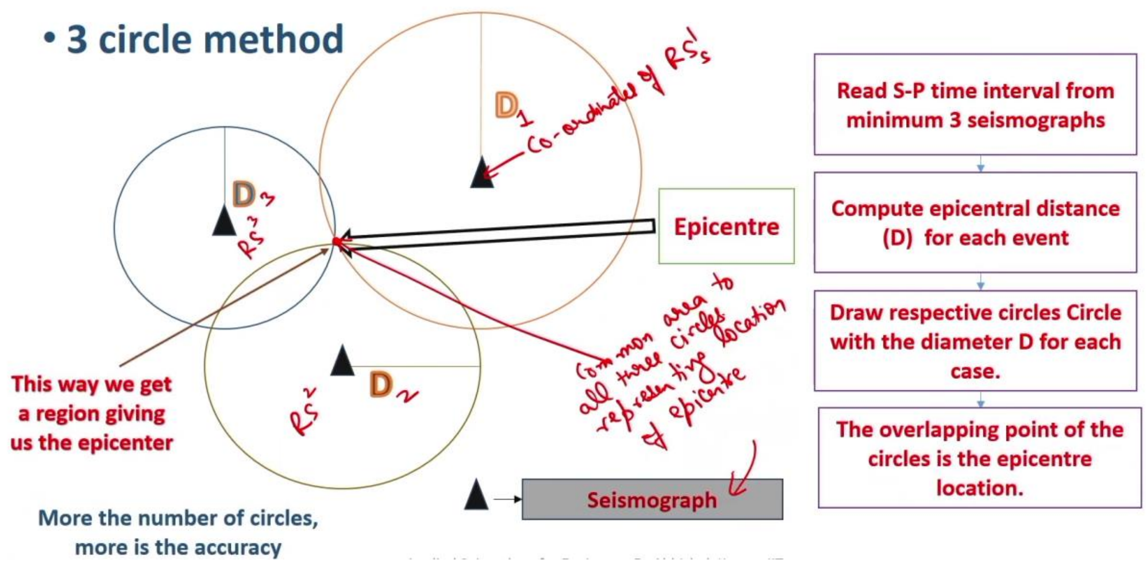
So, as a result, the time of arrival of shear wave will be more in comparison to primary wave arrival time. So, the difference between these two will be always called as T_s minus T_p. Referring to the equations which were given in equation number 1 and equation 2, T_s minus T_p can be correlated with respect to the wave propagation velocity V_p minus V_s over V_p times V_s into capital D, which is the epicentral distance. So, using this, rearranging the terms of equation 3, one can determine the value of capital D as a function of B_p, B_s, T_s, and T_p difference. So, that means, the absolute value of primary and shear wave velocities and the arrival time difference between shear wave and primary wave at a recording station because of one particular earthquake event. So, we cannot take T_s value from one particular earthquake and T_p value from another earthquake or T_s value from one recording station, T_p value from another recording station.

$$\boldsymbol{D = \frac{(V_p \times V_s)}{(V_p - V_s)} \times (T_s - T_p)} \qquad \mathbf{4}$$

So, both these values of T_s and T_p have to be corresponding to the same earthquake and the same recording station; only then can this formula be used. So, the D value, which is the epicentral distance, is correlated with respect to all these parameters. Now the value in order to use equation number 4, the value of V_s and V_p should be known. Unless these values for a particular terrain are known, we cannot use this particular formula. So, one can refer to existing literature to find out how much is the average wave propagation velocity in the medium of interest. So, using this equation, the value of D is an indication of range of distance; it should not be called as range, it is a radial distance with respect to the recording station where the epicenter can be located. So, that means if this is my recording station, considering D as the radial distance, so anything around this particular radius of capital D is the tentative region of the earthquake epicenter. So, this is the tentative region where anywhere within this particular region the earthquake location can be marked. Now, in order to reduce this particular region, which is the probable location for the earthquake epicenter, we have to have similar values of D from more recording stations, and definitely if these recording stations are in nearby regions, we can expect that the circles from multiple recording stations will merge at some common area, which is a possible indication of the epicenter of the earthquake.



So, this is the first method in which the three-circle method where we will take the epicenter, so these are basically representing the coordinates of recording stations, recording station 1, recording station 2, and recording station 3. You will be having latitude and longitude values, and based on these values, you can locate these points, and then based on the value of D1, D2, and D3, calculated using equation number 4, we can develop three circles, which are shown over here.



Now this is the common area which is marked over here, this is the common area to all the three circles. So, this common area can be very small, it can be quite big also. In case it is larger than significant, the very small area, then we can have maybe more recording stations if available that can also be referred so that this particular area, which is the common area to all circles representing epicenter location. So, this location is the seismograph coordinates. Using the record from seismograph one can determine the value of D1, D2, D3, plot these on a graph sheet on the same scale, and then we will be able to find out the location which is common to all the recording stations.

Now, referring to this particular part, there is one numerical where it is given that the arrival time of primary and secondary wave at different seismographs located at different sites means at different locations but certainly in close-by regions.

- **The arrival time of P and S waves in different seismographs located at different sites for an earthquake are listed below**

Seismograph			
Lattitude	Longitude	P-wave Arrival Time	S-wave Arrival Time
$32^o22'30''$	$121^o52'30''$	7:10:11.10	7:10:19.07
$32^o45'30''$	$122^o20'00''$	7:10:06.30	7:10:10.17
$32^o52'30''$	$121^o43'38''$	7:10:9.11	7:10:15.38

It is given that 1 degree of latitude and longitude corresponds to 111.00 km and 88.20 km respectively at $32^o45'30''$. Using the 3 circle method estimate the epicentral location.

So, recording station that P wave arrival time, S wave arrival time is given, and latitude longitude of that particular seismograph is given, and then it is also given that 1 degree of latitude equals to 111 kilometers, and longitude, it is corresponding to 88.2 kilometers respective the given value of latitude. So, generally, when you are converting longitude to kilometers, you will also refer to what is the latitude of that particular location and use it. So, this can, this has to be estimated based on the three-circle method. So, using these values, firstly, we can determine how much is the time difference between P and S waves, and using the latitude value and longitude value, which is also converted to kilometers using the latitude value of that recording station. So, in the end, we will be able to get both the coordinates in kilometers. Using those, we can locate the location of these epicenters on a graph sheet. So, let us see the epicentral distance.

epicentral distance (Station 1)

$$(D_1) = \frac{(T_s - T_p)}{(V_p - V_s)} (V_p \cdot V_s)$$

$$= \frac{7.97}{(6-3)} (6 \times 3) = 47.82 \text{ km}$$

Here, we are directly calling it as epicentral distance for station 1 epicentral distance. So, that can be called as D1 equals to T_s minus T_p, which is already given in the table, over V_p minus V_s times V_p minus V_s. So, if we put the values over here, which are given, we will get T_s minus

T_p value equals to 7.97 seconds, and the average velocity of primary wave propagation and shear wave propagation is taken as 6 and 3 kilometers per second, respectively. So, using this, one can determine the value of this as 47.82 kilometers. Be careful with the units; remember, here the velocity is given in kilometers per second, time is also given in seconds, so the value of D1 should be kilometers.

Epicentral distance (Station 2)

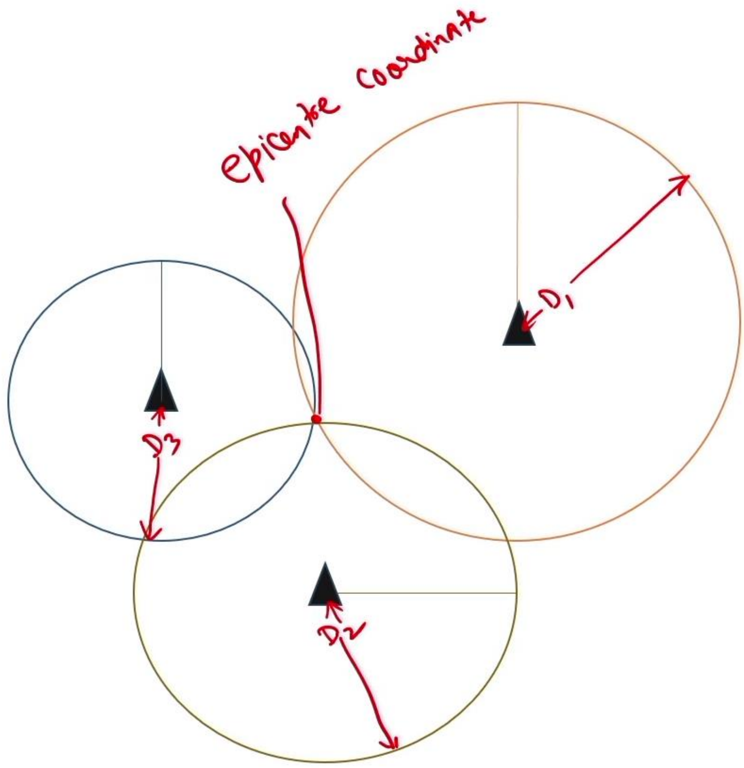
$$(D_2) = \frac{\cancel{(T_s - T_p)} \rightarrow 3.87}{3} \times 6 \times 3 = 23.22 \text{ Km}$$

So, this is the epicentral distance for the second recording station. Distance for station 2, which I am calling as D2, will be equal to T_s minus T_p, which is given over here as 3.87 seconds, divided by 3, which is directly the same value multiplied by 6 into 3, which is taken directly from the above part. This is going to give you 23.22 kilometers as the radial distance from station 2 where the epicenter can be located.

epicentral distance (Station 3)

$$(D_3) = \frac{\cancel{(T_s - T_p)} \rightarrow 6.27}{3} \times 6 \times 3 = 37.62 \text{ Km.}$$

The third part is the epicentral location or epicentral distance with respect to station number 3. So, D3 again can be calculated as T_s minus T_p, which is given for station number 3 as 6.27 seconds, divided by 3 into 6 into 3, which is going to give you 37.62 kilometers. So, we are having the value of D1, D2, D3. Using this and taking the coordinate of the recording station, we will be able to develop these particular 3 circles.



So, I am not actually explaining here how to convert latitude to kilometers and longitude to kilometers. The conversion factor is given in the question itself; you can refer to that and solve to find out the value of coordinates in kilometers, latitude, and longitude for all the 3 recording stations. Once you plot it using a suitable scale on a graph sheet, you will be able to locate point D1, D2, and D3, that is, based on the coordinates of the recording station, then taking the value of D1, D2, D3, which is a representation of this particular distance. D2 is a representation of this particular distance. D3 is a representation of this particular distance. Once we will be able to develop these 3 circles, the common area corresponding to this is basically this particular area. So, on the same graph sheet, one can read how much kilometer along the x-axis, how much kilometer on the y-axis, convert it from kilometer to degrees; one will be able to determine the epicenter coordinate. So, initially we convert it from latitude and longitude values to kilometer values, from degrees to kilimeters. Again, once we are getting from here in kilometers, we will convert that to degrees. So, that is how one can locate the epicenter of the earthquake. If ground motion records, where one is able to locate the arrival time of the primary wave and secondary wave, and B_s value and V_p value of a recording station are known, one can refer to this and determine the location of the earthquake.

Another thing is the intensity of the earthquake. We discuss about whenever there is a wave generated by the source, it is reaching a recording station; we will be able to determine the sense or characteristics of the ground motion, which are detected at a recording station. Now, as we know, recording of a particular earthquake by means of sensors or by means of a rotating drum, where the pen used to mark the signature of the ground motion, has been a recent development. Maybe in the last 40, 50 years, this ground motion recording has become more prominent. However, if we discuss earthquakes and their related damages, most of the damages related to great earthquakes, which have happened in the 1700s, 1800s, 1900s, most of the damages have been known in terms of casualties, in terms of building damage, in terms of maybe some scenario which was witnessed by people living in the epicenter region or in the region where a lot of damages had happened. Certainly, whenever earthquake ground motion records are there, we have those records for understanding the ground motion characteristics and even for generating synthetic ground motions, but at the same time, we cannot completely ignore that such records of ground motion are very limited, which is like for the last 50 or 60 years. However, if we have some information about some damaging earthquake which has happened in the last 200 years, 300 years, like that, we will get more information about what are the damaging characteristics of the earthquake. So, referring to this and in the absence of any ground motion recording instrument, the only way which many times researchers have referred to is measuring the intensity of the earthquake. Please remember the intensity of an earthquake is a qualitative measure. So, one, when is interested in finding out the intensity of the earthquake, it is a qualitative measure, meaning how much characteristics I am able to see at a particular site, what actually I witness which I can call like because of a particular earthquake, this much devastation has happened. That is qualitative because I may say it was very shaky; others may say the shaking was very nominal or moderate, so it is qualitative. But generally, the intensity values are basically a representation of what is the destructiveness of a particular earthquake, how much destruction, how much devastation the earthquake has caused at a particular location. As I mentioned, in the absence of ground motion recording, because that has only started in the last 70 years, 60 years, such information, which is, though not measured by means of an instrument but is giving you qualitative information about the destructiveness of an earthquake, is basically very helpful whenever ground motion record is not there, because it is going to tell you the destructiveness of an earthquake. It is going to tell you how much an earthquake can cause damage, casualties, catastrophes, collapse of the building; even many times, if we refer to intensity maps or intensity scales, it is clearly mentioned the waves are so intense, even it was visible near the ground surface.

So, particle motion was visible near the ground surface was also one of the signatures which people have witnessed, and it cannot be denied that since there was no recording instrument, such ground motion was not witnessed. So, these are very helpful. Generally, it is measured in terms of how the earthquake has affected in terms of ground shaking, in terms of building damage, in terms of damage to dams, bridges, and other structures, as witnessed by a particular person who is staying in that particular area. So, it is like whenever particular damage has happened during a particular earthquake, one can visit those particular sites, as used to happen in earlier days also, that whenever some earthquake caused a lot of devastation, then specific teams or researchers used to visit particular locations. They will interview people and try to understand what actually they witnessed during a particular earthquake. In addition, they will also see if connectivity is resumed within a short span of time; then certainly people can go and witness what damages have happened themselves; otherwise, if the connectivity is not there, then they have to solely rely on the people who are already there in the area under destruction. So, if we go there, we interview people, what actually you have witnessed during a particular earthquake, and depending upon what classification, what information they are going to tell, their standard classification charts. So, we can compare what a particular person observer is telling about the damage of an earthquake; if we are able to witness those damages also by visiting a particular site, provided the connectivity to that particular site in terms of air connectivity, rail connectivity, and road connectivity has been resumed before the rehabilitation work has completely started.

So, once this particular destructiveness of an earthquake is either experienced by an observer or experienced by a researcher or research team, we can compare this with respect to the maps. So, the charts which have been given by different intensity scales, and then, depending upon which particular intensity scale is describing the similar characteristics of devastation, we will assign the same value of intensity to the particular area. Since the values are assigned as per the sense of an individual, there are always uncertainties. I may say it is very damaging; others may say it was moderate, or even the third person who is not that sensible with respect to quick shaking may say I did not feel at all. So, many times there are a lot of uncertainties with respect to assigning intensity to a particular location where the person who is known to the intensity scale has not been exposed directly to the devastations that happened in the particular region.

But as I mentioned that particularly in India, if you refer to earthquakes like the 1905 Kangra earthquake, the 1934 Bihar Nepal earthquake, the 1897 Shillong earthquake, the 1950 Assam earthquake, and many more earthquakes, all these earthquakes caused a lot of devastation, but unfortunately, there were no ground motion recordings available. So, only because of the intensity maps, which are available to us in the present, we are able to understand the damaging characteristics of those earthquakes, the response of important buildings during those earthquakes, and also that will also help in understanding that even this kind of damage, if witnessed in the future, what likely measures can be taken into account so that such damage need not be repeated in the future. It can be minimized to a significant level.

So, qualitative it is; we cannot—we have to keep in mind that it is a qualitative measure, primarily looking into the effects of ground shaking, that too, qualitative effects in terms of damages. Values are assigned as per the sense of an individual; maybe you can interview more than one person and see what they are sensing, and then, based on the classification of damages reported by the people, one can go and assign intensity. So, there are many intensity scales one can refer to. One is the Modified Mercalli Intensity Scale, the Rossi Feral Intensity Scale, the EMS or European Macro-Seismic Scale, the Japanese Meteorological Agency Scale or JMA Scale, and the Medvedev-Spoonheuer-Karnic (MSK) Scale.

So, since the intensity of an earthquake is different, it is not constant. Whenever we are referring to the intensity of a particular earthquake, it is mostly related to the damage that happened at that particular location. So, if I am telling that a particular site has an intensity of 7, that means whatever damage happened at that particular site, X, was a representation of intensity 7 as per a particular intensity scale. Again, I will go to another site, Y, which may be located very close or in a different direction with respect to the epicenter, which might have experienced more damage. Then certainly, I will assign a higher intensity value to that particular location, Y.

Modified Mercalli Intensity Scale

Intensity Level	Description
I	Not Felt
II	Felt only by a few people at rest. Suspended objects may swing.
III	Felt noticeably indoors. Many people do not recognize it as an earthquake. Cars parked may rock a little.
IV	Felt indoors by many, outdoors by few. Dishes, windows, and doors rattle. Parked cars rock noticeably.
V	Felt by most. Many awakened. Some dishes and windows are broken. Unstable objects overturned.
VI	Felt by all. Some heavy furniture moves. Damage slight.
VII	Slight to moderate damage in well-built structures; Considerable damage to poorly built structures; some chimneys broken
VIII	Considerable damage in well-built structures; great damage to poorly built structures; Fall of chimneys, Factory stacks, columns, monuments, walls.
IX	Damage is great in well-built structures, with partial collapse. Building shifted off foundations.
X	Some well-built wooden structures were destroyed; most masonry and frame structures destroyed. Rails bent.
XI	Few of masonry structures stand. Bridges destroyed. Rails bent greatly.
XII	Damage total. Lines of sight and level are distorted. Objects thrown into air.

That means, depending upon the destructiveness, even during the same earthquake, the intensity of damage and the intensity values can be different. So, different values can be assigned to different stations, depending upon how much damage and what the characteristics of damage experienced during the same earthquake at different locations. So, if you join the points having the same value of intensity, you will be able to develop contours. So, contour maps joining the same intensity values are called isoseisms and using the same value, you will have multiple isoseisms because you are going to study the damage in a particular region due to a particular earthquake. So, there will be a number of contours. So, such maps, which are representing a summation of all intensity maps or all intensity contours, are called the isoseismal map of an earthquake. If one is interested, one can refer to what an intensity map looks like. There are intensity maps available for the 1934 earthquake, the 1905 earthquake, the 1897 Shillong earthquake, and many more earthquakes. So, one can refer to the intensity maps which are available for those earthquakes, and even without having ground motion records, that will give an understanding of what the level of damage or devastation was that actually triggered during these historic earthquakes or these earthquakes where ground motion records are not available. So, one particularly, the intensity scale which is widely referred to is the Modified Mercalli Intensity. We can see over here that the intensity values range from 1 to 12. Intensity 1 represents the earthquake was not felt at all; intensity 2, the shaking was felt by a few people, and suspended objects started swinging. In the same way, if you go to intensity 6, felt by everyone, some heavy furniture moved; slight damages were also witnessed during that particular earthquake. Similarly, if you go to 9, damage is great in well-built structures; even well-built structures underwent significant damage; partial collapse also happened in some of the well-built structures; buildings shifted off foundation, so there was tilting in the buildings, and there was differential settlement in the buildings also. Similarly, with an intensity value of 12, total damage—even line of sight and level were distorted, and objects were thrown in the air—so this is basically a more elaborate discussion about when you will assign an intensity value of 12 to a particular recording station or a particular observation site just by comparing the damage which has happened at that particular location with respect to the classification of damage for different intensities given in this particular chart. So, if we are comparing the damage with respect to this particular chart, we will say MMI values. This is called the Modified Mercalli Intensity or MMI. This is generally referred to in terms of MMI, represented in Roman numerals, as mentioned over here. So, depending upon the damage, we'll compare it over here and then assign a Roman numeral of MMI 6, MMI 7, and then later on it will also be used to refer to what kind of damage was witnessed.

Japan Meteorological Agency (JMA) Scale

Intensity	Description
0	Imperceptible to people
1	Some people in the building feel it.
2	Many people in the building feel it. Some people awaken if the quake strikes at night.
3	Felt by most people in the building. Some people are frightened.
4	Many people are frightened. Some people try to escape from danger. Most people awekened if the quake strikes at night.
5 lower	Most people try to escape from danger. Some people find it difficult to move.
5 upper	Many people are frightened and find it difficult to move
6 lower	Difficult to keep standing.
6 upper	Impossible to keep standing and move without crawling.
7	Thrown around by shaking and impossible to move around at will.

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Similarly, the Japan Meteorological Agency also came up with a different scale, where we can see a 0 intensity is also there, and this intensity scale varies from 0 to 7. So, 7 intensity means thrown by shaking and impossible to move around at all. Intensity 6, lower and upper are there so difficult to keep standing; upper is there, impossible to keep standing and without move, without crawling. So, this is basically some description of what is actually witnessed during that particular earthquake shaking, and comparing it with respect to the intensity scale, one can refer to a JMA value of 4, one can refer to a JMA value of 7. So, JMA value of 4, JMA value of 7, so that is basically describing the destructiveness of a particular earthquake at your observation location. As I mentioned, this value keeps on changing as you are moving from one recording station to another site. Many a time, even now also, like the 2015 Nepal earthquake, the Turkey earthquake, even in recent times also, whenever we are getting ground motion records in terms of peak ground acceleration, spectral acceleration values, peak ground velocity, and peak ground displacement values, many a time we also come across the intensity values by agency, so both values—intensity as well as ground motion signature—are reported nowadays.

European Macroseismic Scale (EMS)

Magnitude	Description	Earthquake Effects	Frequency of occurrence
Less than 2.0	Micro	Micro earthquakes, not felt.	Continual
2.0-2.9	Minor	Generally not felt, but recorded.	1,300,000 per year (est.)
3.0-3.9		Often felt, but rarely caused damage	130,000 per year (est.)
4.0-4.9	Light	Noticeable shaking of indoor items, rattling noises, significant damage unlikely	13000 per year (est.)
5.0-5.9	Moderate	Can cause major damage to poorly constructed buildings over small regions. At most slight damage to well-designed buildings.	1319 per year
6.0-6.9	Strong	Can be destructive in areas up to about 160 kilometers across in populated areas.	134 per year
7.0-7.9	Major	Can cause serious damage over large areas	15 per year
8.0-8.9	Great	Can cause serious damage in areas several hundred kilometres across.	1 per year
9.0-9.9		Devastating in areas several thousand kilometers across.	1 per 10 years
10+	Massive	Never recorded, widespread devastation across very large areas; see below for equivalent seismic energy yield	Extremely Rare(unknown/ may not be possible)

Again, the European Macro-Seismic Scale or EMS Scale—here we can see the intensity is correlated with respect to the magnitude also. So, if we are referring to micro, then less than 2 magnitude earthquakes are generally referred to as micro, and then you will say, not felt at all. Similarly, 7 to 7.9, you call those earthquakes major earthquakes, and then cause severe damage over larger areas. If we see in terms of frequency, at least 15 earthquakes per year, on average, are reported across the globe which are major earthquakes. 9 to 9.9 is quite devastating; the devastation can be clearly witnessed in several thousands of kilometers across the epicentral region. Generally, these are reported once in 10 years and even many a time less than that also. Bellow 10, maximum reported earthquake was close to 9.2. Above 10 is not generally reported, so that is why it is not reported so far. So here we can see the classification as well as correlation with respect to magnitude, as well as how frequently such earthquakes are happening across the globe.

Rossi Feral Intensity Scale (1873)

Scale	Description
I	Microseismic shock. Recorded by a single seismograph or by seismographs of the same model, but not by several seismographs of different kinds: the shock felt by an experienced observer.
II	Extremely feeble shock. Recorded by several seismographs of different kinds; felt by a small number of persons at rest
III	Very feeble shock. Felt by several persons at rest; strong enough for the direction or duration to be appreciable.
IV	Feeble shock. Felt by persons in motion, disturbance of movable objects, doors, windows, cracking of ceilings.
V	Shock of moderate intensity. Felt generally by everyone; disturbance of furniture, beds, etc., ringing of some bells.
VI	Fairly strong shock. General awakening of those asleep; general ringing of bells; oscillation of chandeliers; stopping of clocks; visible agitation of trees and shrubs; some startled persons leaving their dwellings.
VII	Strong shock. Overthrow of movable objects, fall of plaster; ringing of church bells. general panic , without damage to buildings.
VIII	Very strong shock. Fall of chimneys; cracks in the walls of buildings.
IX	Extremely strong shock. Partial or total destruction of some buildings,
X	Shock of extreme intensity. Great disaster; ruins; disturbance of the strata, fissures in the ground, rock falls from mountains.

In the same way, with respect to RFI or Rossi Feral Intensity Scale, which was proposed, here we can see again the scale varies from 1 to 10. 1 again refers to recorded by a single seismograph or the shock felt by an experienced observer only. Many a time, people are not sensible with respect to very low values of shaking, but some people are very sensitive, so they will only be able to understand that there has been some shock, some kind of vibration in the ground that has been sensed by an experienced observer. Similarly, if you go with an intensity of 6, fairly strong shock, general awakening of those who were asleep, ringing of bells, oscillation of chandeliers, stopping of clocks; again, with 8, very strong, falling of chimneys, and cracks in the walls of buildings were witnessed. So, if you are witnessing these things, you simply assign an intensity value of 8 on the RFI Scale. Please remember, whenever we are developing an intensity map or whenever we are explaining that a particular site had undergone damage corresponding to that intensity, we have to also refer to which scale one is referring to, because here, going with this particular part, certainly we cannot say RFI value causes RFI of 12 because the intensity scale does not go to 12. So, it goes to 12, but not in RFI; it goes in MMI value. So, one has to be very particular about which particular intensity scale you are using, and corresponding to that value of intensity, whether the actual damage at a particular site was also witnessed or not. So, one has to be very careful while assigning intensity and developing the isoseismal maps.

Now, as I mentioned, firstly, intensity—though it is important because whenever earthquake records were not there, the destructiveness of an earthquake can be correlated very well with respect to intensity values. And it is primarily thus the intensity or isoseismal maps for historic earthquakes which are known, which many a time are referred to even now by generating synthetic ground motion while understanding that such devastation should not be repeated in the near future. But at the same time, it is a qualitative measure. I may say it was very intense; you may say it was moderate or it was minor intensity values. I may say building damage during a particular earthquake; others may say the damage was very minimal to call it as significant damage. So, it is again a qualitative measure, and it certainly depends upon how the person whom we are interviewing, the observer whom we are interviewing to develop the intensity map, how much that particular person is sensible to ground motion. Thirdly, we cannot deny the fact that intensity at a particular station can only be assigned when some signature of ground damage is actually available at the particular site of interest. That means, if there is a site where there is no building, there is no one staying there, certainly how we are going to quantify the intensity of that particular earthquake? So, we cannot assign that intensity even though the ground vibrations were more because no one was actually present there to witness or to feel some characteristics of ground motions or shock, which was witnessed at that particular site during a particular earthquake. So, keeping those in mind, there will always be uncertainty with respect to intensity values because these are based on an individual's experience, which may vary from one person to another. Secondly, the intensity value is not constant; it keeps on changing depending upon the region and how much damage has been triggered in that particular region. So, in order to actually quantify the magnitude of the earthquake, is a better term, which is primarily dependent upon the ground motion signature available at a recording station.

So, unlike intensity, which determines the effect of an earthquake motion on a building or a person, earthquake magnitude is a measure of the size of the earthquake, how big was the earthquake. It is measured; it is a constant value for an earthquake and is not a variable, just like intensity. So, intensity was changing with respect to the site of observation; earthquake magnitude remains constant and is directly related to the size of the earthquake which has happened. So, the size of the earthquake during a particular earthquake is not changing with respect to location; hence the intensity values may change but not the magnitude. Since the earthquake magnitude is determined using suitable ground motion records, the determination is a quantitative value. Intensity was qualitative; this is a quantitative measure, it is going to give you always some value of quantity—how much was the magnitude, and for that particular earthquake magnitude, for that particular earthquake, the magnitude value remains constant. Each step increase in earthquake magnitude represents an increase in amplitude of the vibration by a factor of 10. So, if we are talking about two earthquakes, one is having a magnitude of 5, the other is having a magnitude of 6, then magnitude 6 will have more than 10 times higher amplitude with respect to magnitude 5 earthquake. Thus, vibration caused by magnitude 2 will be 10 times higher than magnitude 1; if we are comparing 3 and 1, it will be 100 times. So, 10 times between 1 and 2 and again 10 times between 2 and 3, so it will be 100 times more. The increase in seismic energy, on the other hand, because whenever a rupture is happening at a particular fault, there will be a release of seismic energy in terms of seismic waves, which are propagating and carrying the seismic energy to larger distances. When this seismic energy is interacting with the medium—soil, structure, building, dams, tunnels—it will sometimes undergo partial damage, sometimes it will be always ground shaking, sometimes increase in pore water pressure, and sometimes complete collapse. So, this increase in seismic energy is 30 times higher when one increases in the magnitude of the earthquake. So, that means magnitude 5 will have will be releasing 30 times more energy in comparison to magnitude 4 earthquake. Magnitude 6 will be releasing 900 times more energy with respect to magnitude 4 earthquake. So, from 4 to 5, 30 times; 5 to 6, 30 times, so 30 into 30 equals 900 times more energy will be released with respect to magnitude 4 when a magnitude 6 earthquake happens.

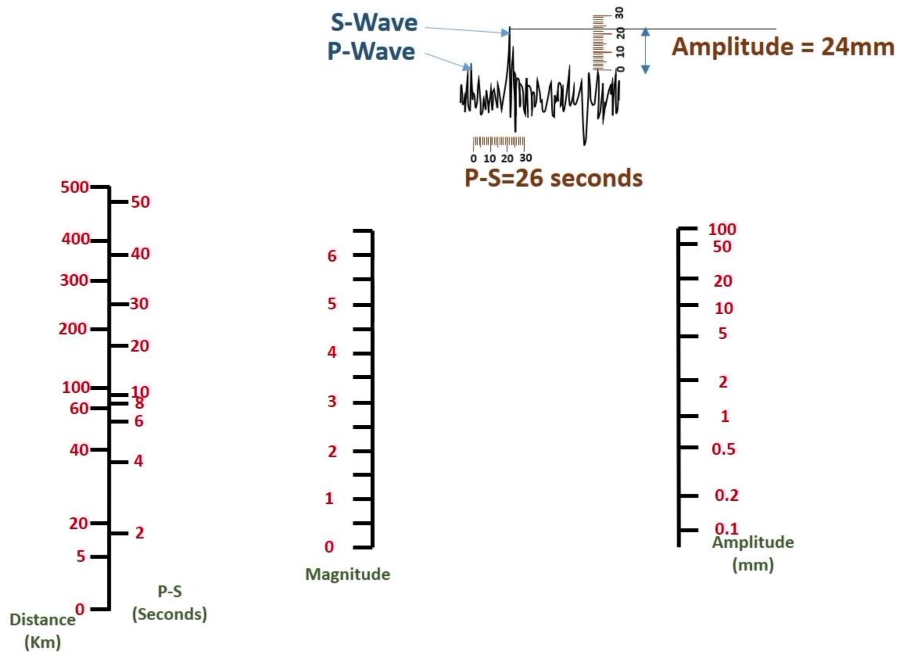
The largest ever recorded earthquake was approximately a 9.2 magnitude earthquake. Beyond 9.2 is not possible many a time because before any further continuation of seismic energy continues in the rocks, there will be some earthquake or there will be some failure in the material alone. So, various magnitude scales have been developed to calibrate the ground motion characteristics and to take those ground motion properties to find out how much energy has been released during a particular earthquake. So, the type of magnitude during a particular earthquake: one is Richter magnitude or local magnitude. As I mentioned, firstly, Richter magnitude came into the picture, which will take into account Wood Anderson seismograph consisting of a rotating drum and marking the ground motion by means of stylus or pencil. This particular scale, many a time, will also be heard about local scales, so more or less, both terms are used back and forth. Another one is seismic moment magnitude or moment magnitude (Mw), which is more permanent with respect to the energy release. Then body wave magnitude, which primarily uses the content of body wave; surface wave magnitude, so these are primarily the four ways in which one will come across whenever there is an earthquake, there is a magnitude of a particular earthquake. In addition, many a time, it is also given that in which particular scale one is referring to those magnitudes—whether it is a body wave magnitude, surface wave magnitude, Richter magnitude, or seismic moment magnitude, or moment magnitude.

So, Richter scale, it was developed, so it was proposed by Charles Richter using a Wood Anderson seismometer. The scale was designed for earthquakes in Southern California and recorded by a network of Wood Anderson seismometers. All the scales which exist today stem from one scale that is Richter magnitude that gives an idea that ground motion signature alone can be used to find out the size of the earthquake.

$$ML=\log(A)+f(\Delta)$$

So, this is now here we can see ML, which is the definition of Richter magnitude. It is a function of log a, which is the maximum amplitude traced in millimeters by a Wood Anderson seismometer. F delta is an empirically determined calibrating function of the epicentral distance delta. The value of delta is 0 for an epicentral distance of 100; for other values of delta, there are calibration charts and empirical correlations also available. So, you put the values of delta over here, then we will get the value of F delta, and corresponding to the same recording station, which is located at delta epicentral distance, how much is the maximum trace amplitude in millimeters. So, using those values of A and F delta put in this particular equation, one will be able to determine how much is the local magnitude or Richter magnitude.

The value of ML obtained at nearby stations is somehow smaller than those of distant stations, primarily because attenuation was not properly accounted for in the Richter scale. It is used to measure the amplitude at a specific frequency, for example, 1 hertz, while the frequency of moderate to large earthquakes beyond will have even lower frequency values. The earthquake size could not be measured from just a single seismometer, thus analyzing data from a multiple number of ground motion records located globally used to delay many a time the establishment of the magnitude of a particular earthquake. And the Richter scale generally saturates at 6.5 magnitude, and thus, many a time whenever a larger magnitude earthquake happens, it will report 6.5 magnitude and thus it will underestimate the seismic energy released during bigger earthquakes.



Now here one we can see about the magnitude of the earthquake taking a ground motion signature. So, here we can see that based on the signature, one we can get is what is the peak amplitude or how much is the peak motion with respect to its mean position which has been detected by a recording station. Using those peak amplitudes and taking the location of the recording station, as well as the difference in the time between P and S wave, one can locate the point on this particular vertical line as well as in this particular line. Joining those two points, whenever it is passing through the magnitude line, this will also give you an understanding about what is the magnitude of the earthquake, which triggered particular P and S wave arrival time difference at a recording station given over here and also triggered a maximum amplitude of vibration as mentioned on the right-hand side line. So, this is independent of the formula which is given earlier. So, using this also, one can determine how much is the magnitude of a particular earthquake.

Now moment magnitude, which is given over here, as all the magnitude scales—let me go to C swing moment magnitude first. So, surface wave magnitude first—let me go through, which was proposed in 1945 by Gutenberg. The Richter scale does not differentiate between different kinds of wave types; that means, whether it is body wave or surface waves, it does not differentiate. At considerable epicentral distance, most of the body wave would have attenuated, and thus most of the damage will be caused by surface waves. So, even at a larger distance, if there is ground motion recorded, how that can be used to find out the magnitude of the earthquake. So, surface wave magnitude is generally determined by taking into account the amplitude of Rayleigh wave marked at a recording station:

$$M_s = log(A) + 1.656 log\Delta + 1.818$$

M_s equals log A plus 1.656 log Δ plus 1.818, where A is the amplitude, combined amplitudes we will be having east-west and north-south components. So, combined amplitude in the east-west and north-south direction, measured in microns, micrometers, I mean, this is the amplitude of Rayleigh wave measured in microns corresponding to 20 seconds, and delta is the epicentral distance in degrees. So, using the value of delta equals to epicentral distance, A is the amplitude at a recording station corresponding to 20 seconds, and the station is located at delta degree epicentral distance. Put those values using this particular equation; one will be able to determine how much is the surface wave magnitude value. The coefficient values many a times change also with respect to the region of interest, but the function formula of the equation remains the same. The equation was primarily developed for Pasadena earthquakes, and thus for different regions, one can have different values. Again, this particular magnitude scale, that is, surface wave magnitude scale, also saturated at 8 magnitude value. Again, body wave magnitude for deep-focus earthquakes, surface wave magnitude will not produce surface waves, will not be produced by deep-focus earthquakes. Hence, reliable estimation of the size of the earthquake will be difficult from body surface wave records. So, one has to go with body wave magnitude.

$$m_b = log(A) + logT + 0.01\Delta + 5.9$$

So, here we can see body wave magnitude equals log A plus log T plus 0.01 Δ plus 5.9. So, here A is the P wave maximum amplitude; maximum is a combination of both horizontal components. Let me say, in the east-west and north-south direction. T is the period corresponding to the maximum amplitude, and delta is the value of epicentral distance in degrees. Put the value of A, T, and delta, and that will give you how much is the value of body wave magnitude during a particular earthquake. Again, this particular scale of body wave magnitude also saturates at 6.5 magnitude. So, we discussed body wave magnitude, and we discussed surface wave magnitude. That means, depending upon the location of the epicentral distance, whenever we are very close, we can refer to body wave magnitude. When you are looking at a distance generally more than twice the thickness of the Earth's crust, more, we can refer to surface wave magnitude. The Richter scale was there, but it was not able to differentiate between surface waves and body waves, and over that, there was saturation with respect to above 6.5 magnitude earthquakes. So, considering the limitations with respect to all these intensity scales, another scale, which was moment magnitude scale, was proposed by Hanks and Kanamori in 1979. So, this particular scale, we can refer to that earlier scales were primarily related to saturation, similarly with respect to the position of the recording station with respect to the location of the earthquake, and some limitations. So, all the earlier scales were used to measure the size of the earthquake based on the amplitude of ground shaking. It was not directly an indication of how much energy was released during a particular earthquake. However, the level of shaking is not directly a function of the earthquake size; with an increase in earthquake size, as we discussed many a times, the amplitude of ground vibration may not increase significantly, primarily at low magnitude, low-frequency values. So, overcoming these limitations of local magnitude scale, body wave magnitude scale, and surface wave magnitude scale, Hanks and Kanamori in 1979 proposed moment magnitude scale, which basically quantifies in terms of how much energy is released during a particular earthquake.

- **Seismic Moment(M_o)**

$$M_o = \mu UA \text{ (Kanamori,1977)}$$

❖ Where μ is shear modulus (G) of the medium (32 GPa for crust and 75 GPa for Mantle) in ($^{Dyne}/_{cm^2}$)
❖ U is the average slip during the rupture (cm)
❖ A is the total area ruptured during the earthquake (cm^2)

So, seismic moment, which is given over here, is a function of mu, which is the shear modulus of the medium undergoing failure. In this particular case, if it is crystal medium where failure is happening, it will be an approximate value; it is given as 32 gigapascals. If it is in the mantle, 75 gigapascals. Certainly, one can refer to the existing literature to find out how much is the shear modulus of a crystal medium as well as in mantle medium to be taken into account to find out the moment magnitude. So, the value of mu in this particular equation should be used in dyne per square centimeter. Similarly, the value of capital U is the average slip that has happened during a particular earthquake and should be measured in centimeters. Capital A is the rupture area, which means the area along the length and width measured on the fault plane in terms of centimeters square. So, using all these terms, mu in dyne per square centimeter, U in centimeters, and A in square centimeters, you can put them over here; then you can determine the value of the seismic moment.

- **Moment Magnitude (M_o)**

$$M_w = \frac{2}{3} log M_o - 10.7 \text{(Hanks and Kanamori, 1979)}$$

Again, that can be converted to moment magnitude as 2 by 3 log M0 minus 10.7, which was again proposed by Hanks and Kanamori in 1979. So, using this, one can determine, firstly, the seismic moment using specific values of shear modulus, slip, and rupture area, and then subsequently, it can be determined with respect to the M_w value, that is, moment magnitude.

- **Seismic Energy "E" (ergs) [Gutenberg and Richter, 1956;Kanamori, 1983]**

$$log E = 11.8 + 1.5 \times M_s \text{ or } M_w$$

Seismic energy also can be determined as log E equals 11.8 plus 1.5 times M_w. So, one can refer to M_w. If existing correlations are there, M_s value also can be converted to M_w, and then you can utilize it. So, this is the method based on which one can determine actually how much seismic energy is being released during a particular earthquake. Since this particular scale is directly related to the amount of energy released during a particular earthquake, this particular scale does not saturate and gives a clear indication of how much energy is released during a particular earthquake. So, let us solve a numerical.

- Problem 1

An earthquake causes an average of 5m strike-slip displacement over a 100 km long, 20 km deep portion of a transform fault. Assuming that the rock along the fault had an average rupture strength of 200 kPa, estimate the seismic moment and moment magnitude of the earthquake.

Solution :

$$M_w = \frac{2}{3} \log M_o - 10.7 \text{(Hanks and Kanamori, 1979)}$$

$$M_o = \mu UA \text{ (Kanamori,1977)}$$

μ= 200kPa=(2×10^6) Dyne/cm^2

A=(20×100) Km^2 = (2×10^{13}) cm^2

U= 5 m =500 cm

An earthquake caused an average slip of 5 meters during strike-slip faulting, and this triggered 100 kilometers and 20 kilometers portion to undergo rupture. Assuming that the rock along the fault has an average rupture strength of 200 kPa, estimate the seismic moment and moment magnitude of the earthquake.

Solution:

So, magnitude is related to seismic moment. Firstly, determine the seismic moment. So, 200 kPa, it should be 2 into 10 raised to the power of 6 dyne per centimeter square. The area which is undergoing rupture or involved in this particular earthquake process is 100 kilometers by 20 kilometers. So, convert the same into centimeters square. U, average slip, which was 5 meters equals to 500 centimeters. Using this, one can determine how much is the seismic moment in terms of dyne-centimeter.

$$\mathbf{M_o} = (2 \times 10^6) \times (2 \times 10^{13}) \times 500 = 2 \times 10^{22} \ \text{dyne} - \text{cm}$$

$$M_w = \frac{2}{3} \log M_o - 10.7$$

$$M_o = 2 \times 10^{22} \ dyne - cm$$

$$M_w = \frac{2}{3} \log(2 \times 10^{22}) - 10.7$$

$$= \frac{2}{3} \times (22.301) - 10.7$$

$$= 14.867 - 10.7$$

$$= 4.1673$$

❑The seismic moment is 2×10^{22} dyne − cm andThe moment Magnitude is 4.1673

Again, using this particular equation, which is given over here, 2 by 3 log M_0 minus 10.7, the seismic moment which is determined over here can be used to find out the value of moment magnitude. So, this is the value one can get.

Earthquake Classification

- Classification based on;
  - **Focal depth**
    - Shallow focus earthquake (<70km)
    - Mid focus or intermediate depth earthquake (70 km to 300 km)
    - Deep focus earthquake (300 km- 700 km)

    [No earthquake occur > 700 km focal depth]
  - **Magnitude**
    - 3-3.9 **Minor EQ**
    - 4-4.9 **Light EQ**
    - 5-5.9 **Moderate EQ**
    - 6-6.9 **Strong EQ**
    - 7-7.9 **Major EQ**
    - ≥8.0 **Great EQ**
  - **Based on Location**
    - Interplate (Occur at the Plate Boundary)
    - Intraplate (Occur within a Plate)
  - **Based on cause**
    - Tectonic EQ (during to slip along casaultive fault)
    - Non-tectonic EQ (associated with Volcanoes or manmade reasons such as blast, construction of dams, blasts
  - **Based on Epicentral distance**
    - Local EQ (within 1 degree epicentral distance)
    - Regional EQ (1 to 10 degree of the epicenter)
    - Teleseismic EQ (greater than 10 degree)

Now, we discuss intensity, we discuss the magnitude. Using the information about earthquake parameters, we can classify the same earthquake based on focal depth, based upon the range of epicentral distance, based upon where it has happened, and based upon the cause of the earthquake. So, in a variety of ways, one can classify the earthquake. The first one is given based on the focal depth. So, if the focal depth of the earthquake is less than 70 kilometers, you call it as shallow-focus earthquake. If it is between 70 to 300 kilometers, you call it as mid or intermediate focus earthquake. If it is between 300 to 700 kilometers, you call it as deep-focus earthquake. Generally, no earthquake is reported beyond 700 kilometers focal depth. Similarly, with respect to magnitude, if the magnitude of the earthquake is 3 to 3.9, it is called a minor earthquake, and subsequently, all classifications are given over here. Many a times, we will come across a term called "great earthquake." So, "great earthquake" means not in terms of damages; it is certainly a terminology given to an earthquake which is having a magnitude of 8 and above. So, anything which has a magnitude of 8 and above is called a great earthquake. In terms of location, if it is happening at the plate boundary, one can call it an interplate earthquake. If it is within the plate boundary, within the plate, but away from the plate boundary, it is called an intraplate earthquake. Based on the cause, one can call it as tectonic earthquake if it is based on seismic activity. If it is not based on seismic activity, one can call it as non-tectonic earthquake. An earthquake, the same earthquake, based on the focal depth, based on the magnitude, based on the location, based upon the cause, can be classified as a shallow-focus earthquake, minor earthquake, intraplate earthquake, non-tectonic earthquake. If the location is within 1 degree of epicentral distance from the recording station, you can call it as local earthquake. If it is between 1 to 10 degrees, you call it as regional earthquake. If it is above 10 degrees, you call it as teleseismic event. So, these all we have discussed already.

Magnitude	Intensity
• Quantitative	• Qualitative
• Size of earthquake and is based on earthquake records at seismic stations	• Based on the amount of damages and is based on personnel judgment
• A single value for one event	• Vary from place to place for same event.
• Independent of soil and surface condition.	• Depends upon the soil and surface condition.
• Values reported in Arabic numerals till one decimal of place.	• Values reported in Roman numerals up to one decimal place.
• Mathematically no upper limit of measurement	• The range of intensity as per various scales is fixed.
• Values are a function of seismic energy released at the source.	• Related to the effect of the ground surface.

Now, magnitude is qualitative; quantitative intensity is qualitative. Magnitude is measured in terms of energy release; intensity is based on the amount of damages reported. Magnitude is generally reported up to the first decimal place; intensity is reported generally in roman numbers. So, it is independent of size and surface condition, but magnitude certainly depends upon surface condition because that will govern or control the surface scenario of damage. So, one can differentiate between magnitude and intensity. It is like the amount of energy which is required to switch on one bulb remains constant: so, that is an indication of magnitude, and the intensity of light if you are very close to the bulb, and as you move away from the bulb, the intensity of light keeps on changing. So, one can say how the intensity of light is changing with respect to distance; the same way with respect to earthquake intensity, depending upon the damage, one can say the intensity value is changing maybe away from the epicenter or even within the epicentral distance.

The last term for this particular lecture is seismic wave attenuation. Generally, whenever waves are generated at the source and start propagating, directly they will not reach a recording station between the source, which is located maybe a certain kilometer depth beneath the ground surface; the wave will start propagating, and when these waves are propagating through the medium, these waves will cause particle oscillation. These waves, as they move from the epicenter, are moving in three-dimensional space. So, again there will be a lot of scattering happening over here collectively because of these processes, whether it is heat because of particle motion, because of scattering, or inelastic attenuation; there will be redistribution of energy at every point the wave is progressing in the propagation medium. As a result of this redistribution, generally a decrease in the amplitude of the wave away from the focus is observed. So, this phenomenon, which is resulting in the attenuation or decrease in the amplitude of the wave, is called as seismic wave attenuation. This primarily happens due to two factors: one is geometric spreading because, as you are moving away from the epicenter, a larger area, because the wave is distributing in three-dimensional space. So, as you move away from the epicenter, though it is starting from a point or a particular rupture area, as it grows, a larger area is now involved. So, there will be redistribution of energy; it is covering a larger area.

So, geometric spreading, larger geometry is now involved in which the waves are spread. Spherical wave fronts, that is what I am mentioning. The geometrical spreading accounts for the reduction in the amplitude of a given seismic wave front as the area of the wave front increases, and one has to conserve the energy. So, as you move away, there will be a reduction in wave energy.

The second one is inelastic attenuation. So, when waves are interacting with the medium, we have discussed that each time every wave is propagating through a particular medium, there will be oscillation in the particle; because of this motion, there will be relative motion in the particle. Many a time, there will be the generation of heat. So, because of this heat, again, further, there is a reduction in the energy the wave was carrying to larger distances away from the epicenter. So, inelastic attenuation again can lead to a reduction in the amplitude, again can lead to attenuation in the seismic waves as you move away from the epicenter. Inelastic attenuation can be quantified with respect to the quality factor, which is proportional to the ratio of mean energy contained in one cycle to the energy dissipated during one cycle. In addition to this, if along the propagation medium there is heterogeneity present, again that heterogeneity will cause redistribution of energy that will have an effect on the amplitude of the wave.

So, collectively, when we are discussing geometric spreading, because of medium heterogeneity, because of heat, because of inelastic attenuation, collectively all these are leading to an increase in the amplitude of the wave, which is reducing as you are moving away from your focus. So, thank you all. With this, we have come to the end of lecture 8, which gives a broader perspective about how one can determine the magnitude of an earthquake, how one can quantify the intensity of the earthquake, how one can locate a particular earthquake, and whenever waves are passing through a particular medium, what are the different phenomena happening leading to the reduction in the amplitude of the wave. So, thank you, everyone.