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Engineering Seismology

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1. Engineering Seismology

Engineering Seismology is the study of Seismology as related to Engineering. This involves understanding the source, the size and the mechanisms of earthquakes, how the ground motion propagates from the source to the site of engineering importance, the characteristics of ground motion at the site and how the ground motion is evaluated for engineering design. This subject is therefore related to the hazard of earthquakes. The seismic hazard at a site cannot be controlled. It can only be assessed. In the same context, **Earthquake Engineering** is the subject of analysis and design of structures to resist stresses caused by the earthquake ground motion. Resisting the stresses imply either resisting without failure or yielding to the stresses gracefully without collapse. This subject is related to the vulnerability of built structures to seismic ground motion. The vulnerability is controlled by design. The decision to control the vulnerability of a structure is based on the economics of the situation and on the judgement about the acceptable risk to the community. See figure 1.

Therefore, the assessment of seismic risk is based on the seismic hazard, the vulnerability and the value of the loss. This is expressed by the relation:

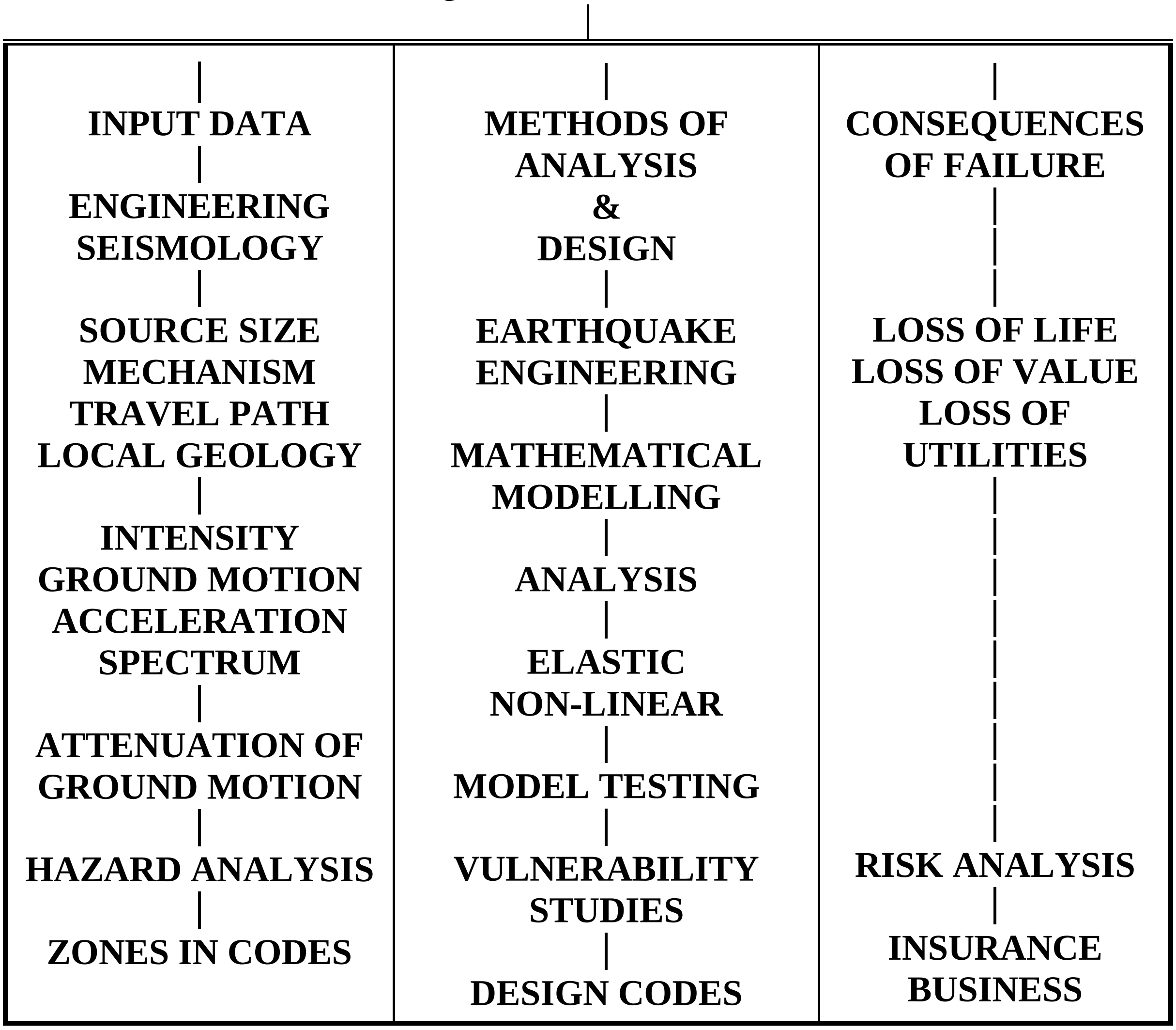
Risk = Hazard * Vulnerability * Value.

The value may be taken in the sense of cost of replacement and is really the problem of insurance business. Ref: Fournier d'Albe (1982).

In this context, "Seismic Hazard" is defined as the probability of occurrence of a ground motion of a given size within a given period of time at the site of interest. This will depend on the possible sources of earthquakes within a reasonable distance of the site and the seismic activity of these sources in relation to size and time.

The "Vulnerability" is a measure of the probability of damage (loss) to the structure to a ground motion of a given size. Different structures have different vulnerability curves. Figure 2 expresses the concepts schematically.

ENGINEERING SEISMOLOGY & EARTHQUAKE ENGINEERING



RISK = HAZARD * VULNERABILITY * VALUE

Figure 1

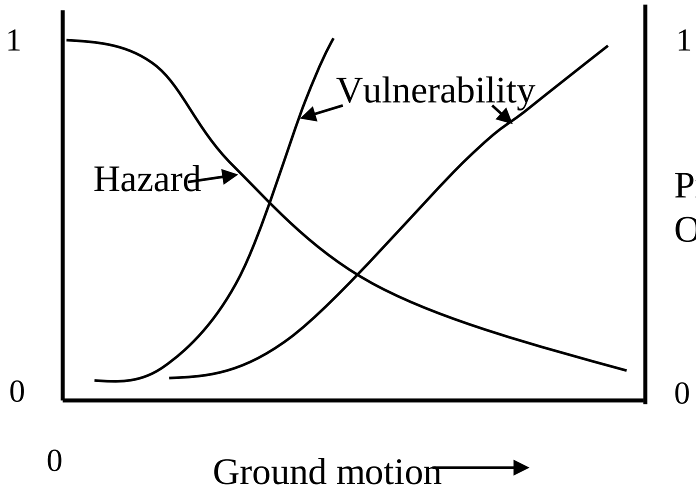


Figure 2

Seismicity and Plate tectonics:

If we look at a map of seismic activity around the world, Figure 3, we notice that earthquakes predominantly happen along some belts. For example the Circum-Pacific belt (around the pacific ocean) or the Alpide belt starting at far east and follow all the way to Europe. However, we note that earthquakes do happen infrequently elsewhere. There is no place on the globe which can truly be said to be non-seismic.

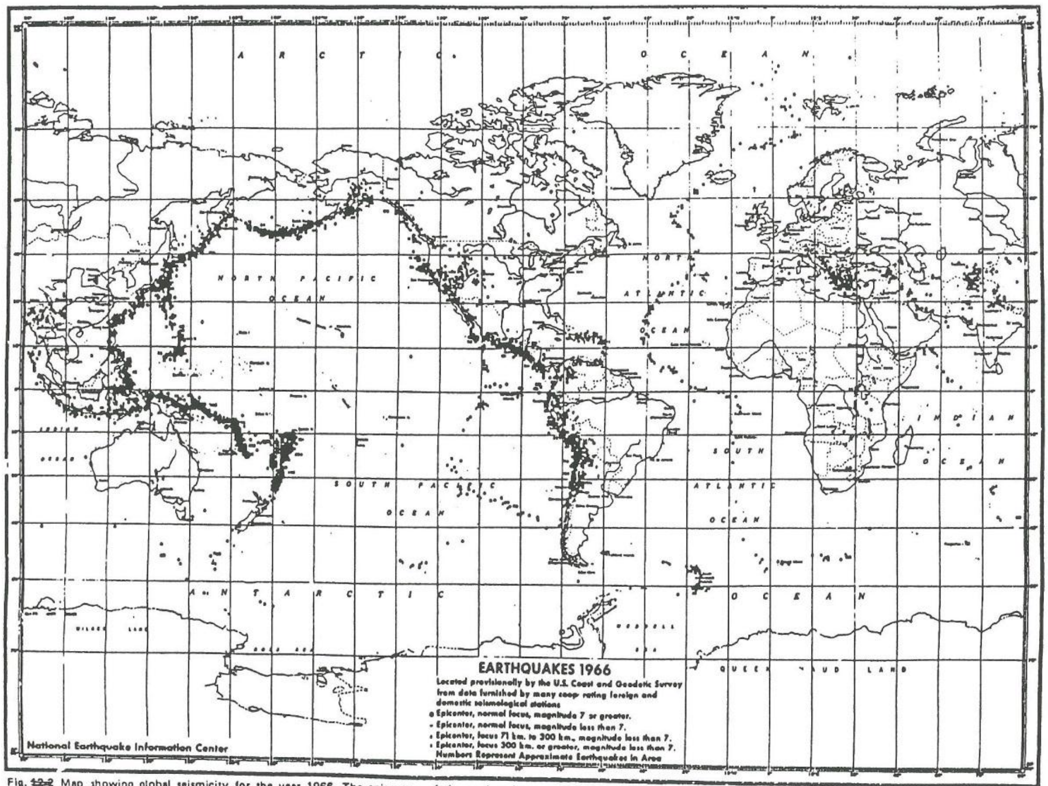


Fig. ~~12-2~~ Map showing global seismicity for the year 1966. The epicenters of the earthquakes were calculated by the U.S. Coast and Geodetic Survey from the observed travel times of seismic P-waves to the seismographic stations. The map can be considered almost complete for earthquakes with a magnitude above 4.[12-2]

Figure 3

The fact of the seismic belts and other reasons led to the Plate Tectonic theory. [Continental Drift theory is a precursor to the Plate Tectonic Theory when Wegner in 1915 showed the global jigsaw puzzle of fitting continents together.] According to this theory, the crust of the earth (more accurately the lithosphere which includes the crust and a small part of the upper mantle) is broken up into about 20 rigid plates, figure 4.

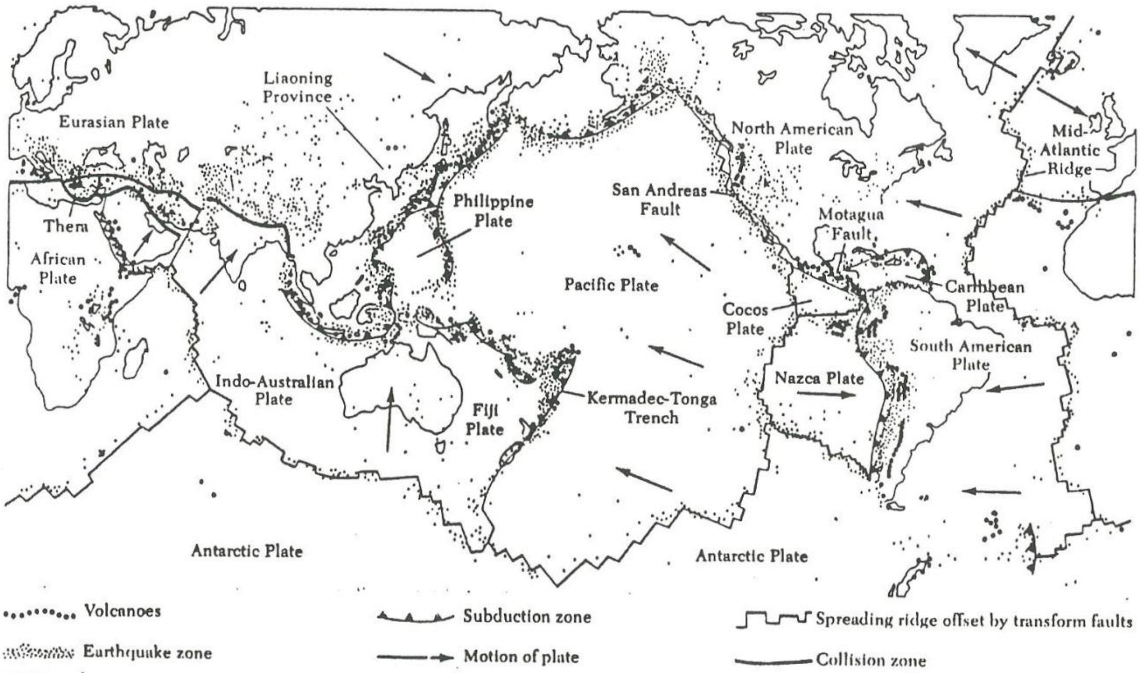


FIGURE 4
World map showing relation between the major tectonic plates and recent earthquakes and volcanoes. Earthquake epicenters are denoted by the small dots, and the volcanoes by large dots.
(from: Earthquakes-A Primer by B.Bolt,1978)

Figure 4

These plates are slowly moving relative to each other continuously. Sometimes, these movements are arrested by static friction and stresses build up. When the stress is sufficiently large it overcomes the friction and the plates suddenly move apart. The model of a conveyor belt in contact with a spring-mass system perhaps explains this mechanism, figure 5. This explains the reason behind the earthquakes along the boundaries of the plates. However, the plates are not entirely rigid and the boundary stresses may cause failure inside the plate as well. That is why, no part of the globe is safe against earthquakes.

What causes the plates to move is only a hypothesis- the heat loss in the earth's interior causing convection current to develop in the mantle and the current carries the plates with it.

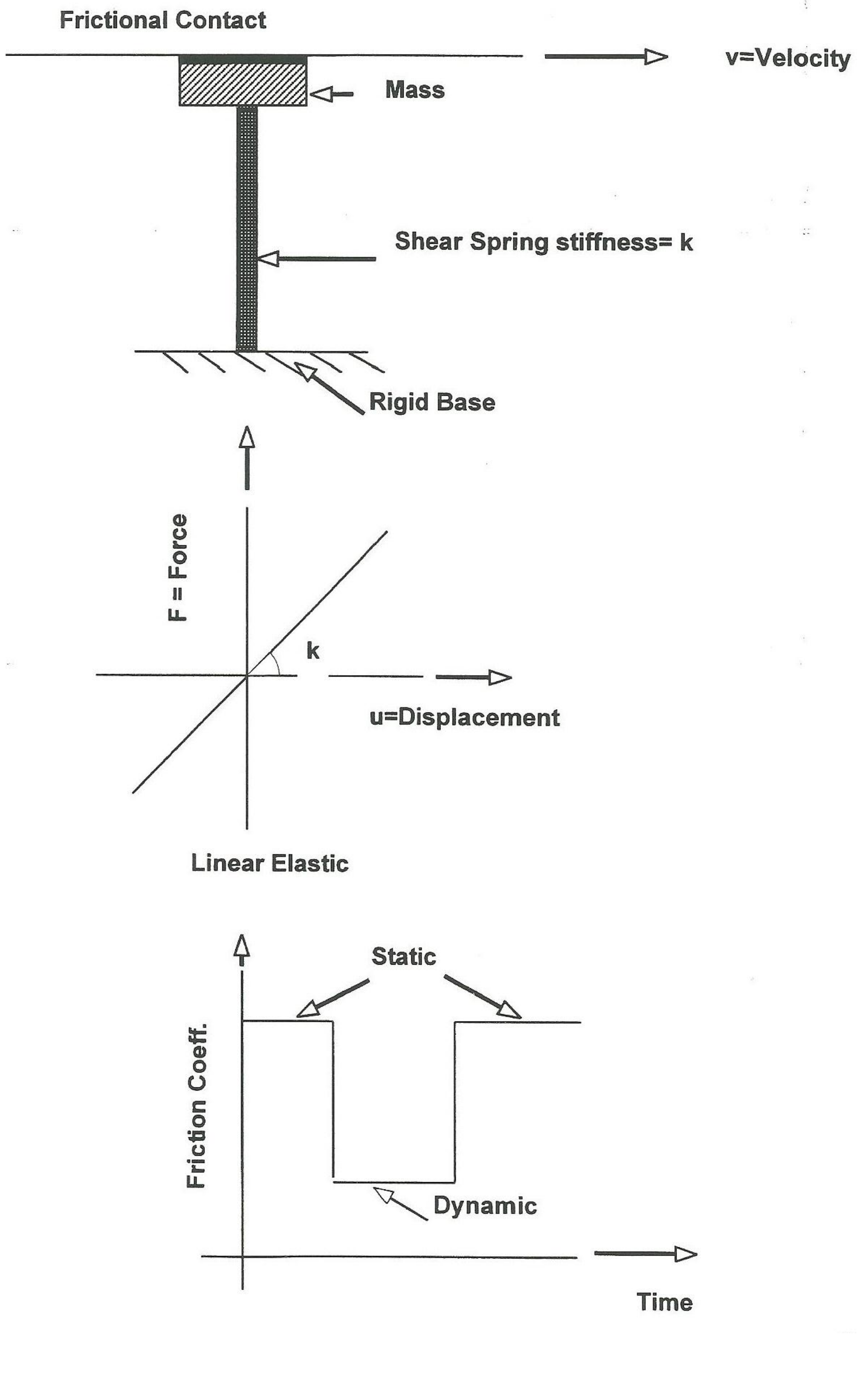

Frictional Contact
v=Velocity
Mass
Shear Spring stiffness= k
Rigid Base
F = Force
k
u=Displacement
Linear Elastic
Static
Friction Coeff.
Dynamic
Time

Figure 5

Figure 6 shows the make up of the earth's interior. This shows schematically the plate interactions. This shows the divergent mechanism creating new plates along the oceanic ridges. The convergent mechanism shows the subduction of one plate under another (convergence of an oceanic plate and a continental plate will cause the dense oceanic plate to slide under the less dense continental plate) or creating mountain ranges (when two continental plates converge).

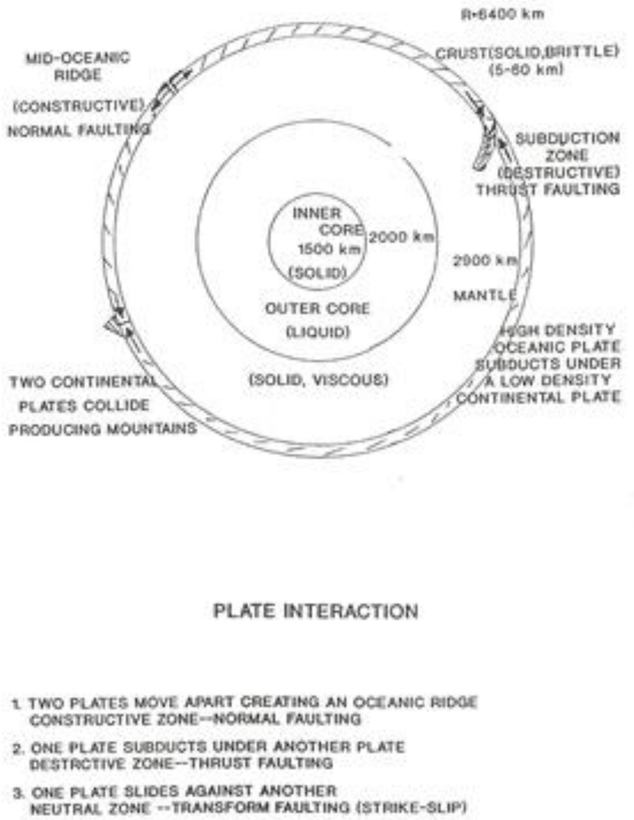


Figure 6

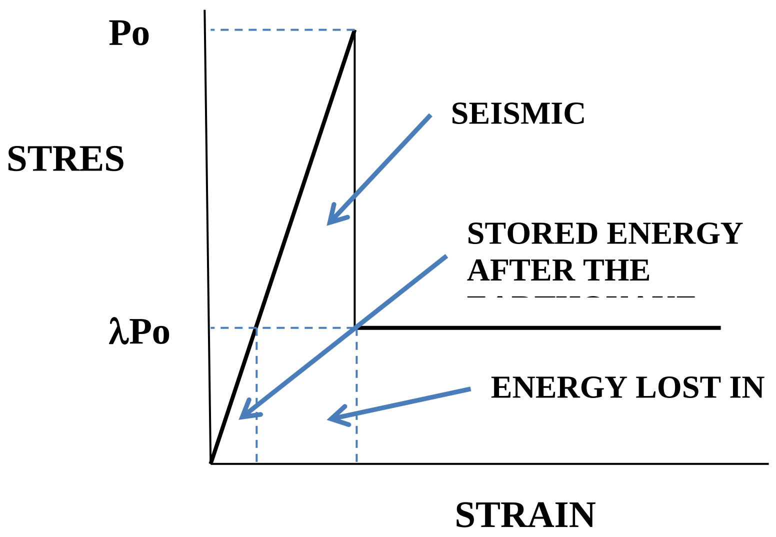
The Source Parameters

Faults and Energy Release:

Earthquakes are the vibrations of ground caused by the sudden release of strain energy stored in the earth's crust. Figure 7 shows the schematic stress-strain curve of the crustal material and the part of the stored energy which is released as an earthquake.This energy is released by the brittle failure on faults and is carried away by the propagation of seismic waves. Since brittleness of the crustal material is an essential part for the sudden release of energy, earthquakes can happen only in the upper part of the earth's crust. Most earthquakes, particularly the damaging ones, are of shallow origin. The deeper earthquakes happen in the subducted part of the crust before it melts in the heat of the mantle. The earthquakes give rise to two kinds of ground movements- a permanent displacement at the fault and its vicinity and the transient ground motions resulting from the propagation of seismic waves away from the source.

The propagation of seismic waves away from the source.

<u>EARTHQUAKE</u> IS THE VIBRATION OF THE EARTH CAUSED BY THE <u>SUDDEN RELEASE OF ENERGY</u> IN A STRESSED VOLUME OF EARTH



SEISMIC ENERGY IS THE PART OF ENERGY THAT IS RELEASED IN THE FORM OF SEISMIC WAVES

ENERGY BEFORE THE EARTHQUAKE / VOLUME $= \frac{1}{2} P_0^2 / G$

P_0 = FAILURE STRESS

G = SHEAR MODULUS

λ = STRESS DROP FACTOR

$$\frac{\text{SEISMIC ENERGY}}{\text{ENERGY BEFORE EARTHQUAKE}} = (1 - \lambda^2)$$

ENERGY/VOLUME = CONSTANT = (0.10 to 0.15) (kN/m^2)

Bigger Earthquake = Bigger Volume = Larger area of Fault = Longer Length of Fault

Figure 7

Elastic strain energy builds up on a fault, which is held static by friction, until the stresses overcome the strength and slip is initiated. Since nature favours an existing fault (finds it easier to break) than a new one, the same faults move repeatedly in successive earthquakes. This does not mean that new faults cannot ever be generated and therefore, theoretically, no part on earth is ever safe from earthquakes

There are three basic types of fault movements, figure 8. These are normal, thrust and strike slip movements. These involve extension, shortening and lateral movement of the earth's crust respectively. Within a small geological time scale, the type of motion in a fault is observed to be the same in different earthquakes thus creating geomorphological features which can be identified.

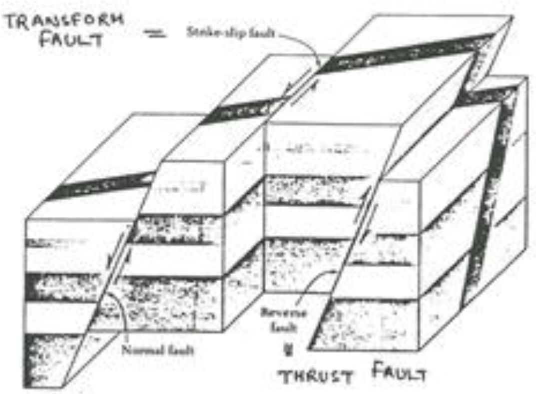


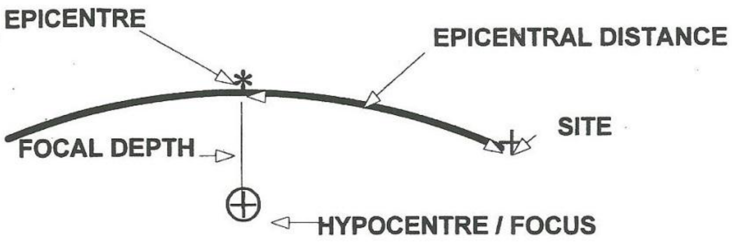
FIGURE 8
Diagram showing the three main types of fault motion.

(From: Earthquakes-A Primer by B.Bolt,1978)

Figure 8

Some definitions:

Here are some definitions, which are supposed to be common knowledge but has some implications. See figure 9.



EPICENTRE

THE POINT ON THE SURFACE OF THE EARTH, BELOW WHICH THE EARTHQUAKE ORIGINATES. IT IS ALSO THE AREA WHERE MAXIMUM DAMAGE OCCURS.

INSTRUMENTAL EPICENTRE ---- POINT (LATITUDE, LONGITUDE)

MACROSEISMIC EPICENTRE ---- AREA OF MAXIMUM DAMAGE

HYPOCENTRE/FOCUS

THE SOURCE OF THE EARTHQUAKE BELOW THE EPICENTRE

INSTRUMENTAL HYPOCENRE -POINT(EPICENTRE, FOCAL DEPTH)

Figure 9

Hypocentre or Focus:

The point on the fault where slip is first originated. From this point, the slip propagates and spreads over the rupture surface (the fault) until the slip is stopped by either strong material or less stress. The hypocentre is represented by three coordinates: Latitude, Longitude and the depth from the earth's surface. Note that the whole fault does not move at the same instant.

Epicentre:

The point on the earth's surface immediately above the hypocentre. It is represented by the latitude and longitude of the point. The error in the determination of the epicentre is about 10km presently. But in the old days, this error could be very large. There are instances of the determination in the wrong hemisphere. It is therefore essential to correlate the instrumental determination of epicentre with the area of maximum damage.

Focal depth:

This is the depth of focus below the epicentre. There are three grades of depth- Shallow, Intermediate and Deep. Most continental earthquakes are shallow and these are of engineering importance. Focal depth of an earthquake is the most difficult one to determine and should be treated with caution. In the bulletin of earthquakes, most earthquakes are given a focal depth of 33km which simply imply that these are of shallow depth but the depth was not possible to determine any more accurately.

Size of earthquakes:

The magnitude and the moment of an earthquake measure the size of an earthquake.

Magnitude:

The magnitude is derived from instrumental readings of ground displacements. These are empirically related to the energy of the earthquake at source and are in logarithmic scale. Magnitude is derived from the amplitude of ground movements at particular frequencies and then correcting it for distance of the source to the recording site. Measurements are made at about 20 sec period to give surface wave magnitude (M_s) or at about 1 sec period to give the body eave magnitude m_b. Local magnitude M_L (commonly known as Richter magnitude) was originally defined by Richter in 1935, Richter (1958). This is the logarithm of the maximum amplitude (recorded on a Wood-Anderson Seismograph in mm) and corrected for distance of the recording site from the epicentre.

$M_s = \log (A/T)_{max} + 1.66 \log (\Delta^o) + 3.3$, known as the Prague Formula,
$m_b = \log (A/T)_{max} + Q (h, \Delta^o)$; Q values are tabulated in literature
$A = A_t/V \bar{M}$
A_t is the measured amplitude in microns, V is the static magnification, $\bar{M}$ is the dynamic magnification.

$M_L = \log A - \log A_0$

A is the maximum amplitude in mm in a standard Wood Anderson Seismograph (T_o=Perod=0.8sec, V=Static magnification=2800, λ=Damping=0.8)
A_0 is the distance correction factor.
[This definition is equivalent to saying that M_L is the logarithm of the measured amplitude in microns at 100 km.]

From the tabulated values given by Richter(1958), log A_0 can be expressed as: (Note: The curve fitting is done by the author and not by Richter)

$\log A_0 = -1.3892 - .0028D - .0007D^2$; $D \leq 35$ km
$\log A_0 = -.78747 - .00272D - .96201 \log(D)$; $D > 35$ km
At D=100km, $\log A_0 = -3$.

Other attempts to quantify the size of an earthquake are by the amount of damage to man-made structures at the epicentral region and by the farthest distance at which the earthquake is felt. There are empirical relationships connecting these parameters to the magnitude. Also empirical relationships exist for connecting the magnitude with the length of faulting etc., see figure 11. However, visible faults may not be there for all earthquakes.

There is another magnitude, called Moment Magnitude, M_w, (Now a days this magnitude is denoted by **M**) which is now being used as the most reliable measure of energy. This is derived from another measure of the size of the earthquake called the Seismic Moment.

The relationship between the magnitude and the energy of the earthquake is empirical, figure 12. As long as the wavelength at which the earthquake is measured (roughly 80km for M_s) is long compared to the length of fault, the logarithmic nature of seismic energy with magnitude is good. When the length of fault is longer than the wave length, the instrument does not see the wave clearly and the magnitude saturates. Figure 13. The slope of the magnitude versus energy curve starts to flatten. Very large energy release is then not represented by the M_s. For m_b, the flatttening happens much earlier.

SIZE OF AN EARTHQUAKE

THE SIZE OF AN EARTHQUAKE REFLECTS THE RELEASED ENERGY. THIS IS MEASURED BY THE FOLLOWING:

1) MAGNITUDE

- **a) BODY WAVE MAGNITUDE** = m_b
- **b) SURFACE WAVE MAGNITUDE** = M_S
- **c) LOCAL MAGNITUDE** = M_L
- **d) MOMENT MAGNITUDE** = M_w

2) MOMENT=--- = M_0

MAGNITUDE AND MOMENT ARE DERIVED FROM INSTRUMENTAL READINGS.

SIZE CAN ALSO BE INFERRED FROM THE FOLLOWING:

3) MACROSEISMIC INTENSITY

4) FELT RADIUS

DEFINITIONS:

$m_b = \log (A/T)_{max} + Q(h,\Delta^0) \Rightarrow T \cong 1$ second

$M_s = \log (A/T)_{max} + 1.66 \log \Delta^0 + 3.3 \Rightarrow T \cong 20$ seconds

$A = A_t / VM$

A_t = measured amplitude in microns
V = Static magnification
M = Dynamic magnification

$M_L = \log (A_t)_{max} - \log (A_0) \Rightarrow$ STANDARD WOOD-ANDERSON SEISMOMETER

$T_0 = 0.8$ second, $V = 2800$, $\lambda = 0.8$

Figure 10

ASSESSMENT OF MAGNITUDE FROM FIELD EVIDENCE

$M_s = 5.86 + 0.4 \log (L^{1.58}\, \delta^2)$

L (km) = Length of fault,
δ (m) = displacement across fault.

Modified from Ambraseys & Melville (1982) – obtained from 63 observations

$\log (L) = 0.7\, M_s – 3.24$,

Ambraseys & Melville (1982) – obtained from 220 observations

Similar relationship can be derived from the definition of Moment (defined later)

$\log (L) = 0.5\, [\, c + d\, M_s – \log (eGH)]$

L = Length of fault, H = Depth of fault, G= Shear Modulus, δ= fault displacement, $e=\delta/L$ --- All have consistent dimensions.

$\log (M_0) = c + d.M_s$ --- M_0 has consistent dimensions with the above

for M_0 in dyne.cm , c=16, d=1.5 and $G = 3x10^{11}$ dynes/cm^2 , $e \approx 1 – 6\ (x10^{-5})$ and converting L from cm to Km and assuming H = 10 km = 10^6cm

$\log (L_{km}) = 0.75\, M_s – 3.25$ **-- (comparable to observed values)**

dyne = gm.cm/sec^2 , 1 N = 10^5 dyne , 1 dyne/cm^2 = 0.1 N/m^2

Ref: Ambraseys N.& Melville C. (1982)- A history of persian earthquakes, Cambridge University Press.

Figure 11

ENERGY OF EARTHQUAKES

$\log E_{ergs} = 11.8 + 1.5\ M_s$ --------Gutenberg & Richter (1956)

$\log E_{ergs} = 12.24 + 1.44\ M_s$ --------Bath (1966)

M_s	E (ergs)	V (cm^3)	L(km)
0	$1.7x10^{12}$	$1.7x10^{9}$	$5.7x10^{-4}$
3	$3.6x10^{16}$	$3.6x10^{13}$	$7.2x10^{-2}$
5	$2.7x10^{19}$	$2.7x10^{16}$	1.8
6	$7.6x10^{20}$	$7.6x10^{17}$	9.1
7	$2.1x10^{22}$	$2.1x10^{19}$	46
8	$5.7x10^{23}$	$5.7x10^{20}$	229

Assumes

$E/V = 1000(dynes/cm^2)$

$\log (L_{km}) = 0.7M_s - 3.24$

Gutenberg N. & Richter C.(1956): Magnitude and energy of earthquakes. Annali di Geofisica, 9, 1-15

Bath M.(1966): Earthquake energy and magnitude. Physics and Chemistry of the Earth. 7, 115-165.

Figure 12

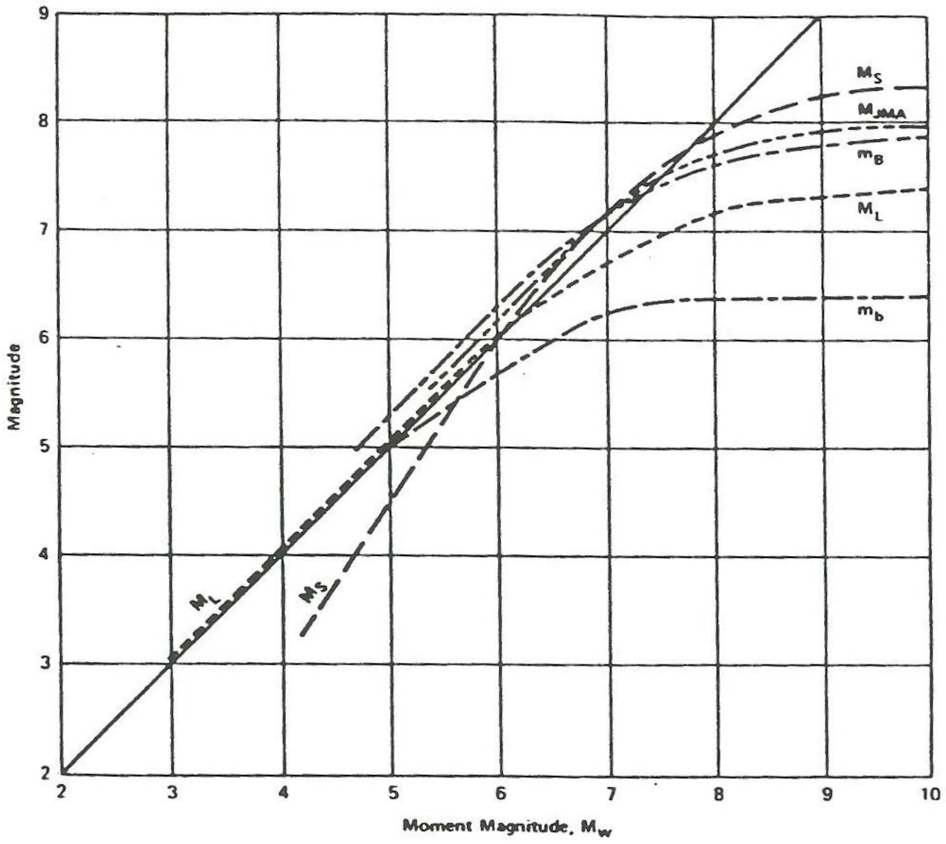


Figure 13: Relation between moment magnitude and various magnitude scales: M_L = Local magnitude, M_s = Surface wave, m_b = short period body wave, m_B = Long period body wave, M_{JMA}= Japan meterological Agency.
(After Heaton et al, 1982, reproduced in Idriss 1985)

Seismic Moment:

Seismic moment is defined as the following: See figure 14.

$M_0 = \mu A \mathbf{u}$

μ is the shear modulus of the medium ($\mu \approx 3x10^{10} N/m^2$), A is the fault area ($m^2$) and **u** is the vector displacement (m) of one side of the fault relative to the other. M_0 has the dimension of (Nm).
M_0 can be calculated from direct measurement in the field if available. This can also be measured from the long period level of the seismic spectrum. Observations from large earthquake show that the fault displacement has a consistent ratio to the fault length ($1\text{-}6 \times 10^{-5}$).

There are relationships linking the moment M_0 to the magnitude M_s . Hanks & Kanamori (1979) gives:

$\log M_0 = 1.5 M_s + 16.1$; In this formula, M_0 is in dyne.cm . [$1Nm=10^7$ dyne.cm]

[Note that the inversion of this formula defines M_w]

The linear relationship between log M_0 and M_s does not seem to be true for smaller magnitudes. Other non-linear relationships exist, for example, Ekstrom & Dziewonski (1988) and Ambraseys & Free (1997).

We can use these relationships to estimate the maximum possible magnitude in a fault or estimate the permanent fault displacements in a major earthquake in the fault. For example: If a capable fault exists which is say 200 km long and 10 km deep (Anatolian fault for example), the estimated maximum fault displacement can be of the order of $5x10^{-5}L$ which will be about 10m. The moment M_0 for this earthquake will be:

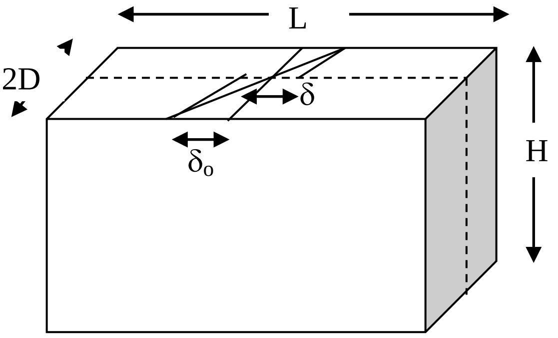
$M_0 = 3.3x10^{10}(200x10x10^6)x10$ Nm$=6.6x10^{20}$ Nm $=6.6x10^{27}$dyne.cm.

This will then convert to $M_s = 7.8$. Thus, on a fault of the size of 200km, 7.8 magnitude earthquake may be expected.

Similarly, we may estimate the fault displacements for earthquakes of various magnitudes. We can see that displacements across faults for a medium size earthquake, say of magnitude M_s=6 may be of the order of a meter. Using $\log(L_{km})=0.7M_s -3.24$ will give $\mathbf{u} = 0.47$m. Thus, if we are considering a dam across a fault and the design earthquake is a magnitude 6 one, then the design must consider a possible fault movement of 1/2 metre. Note that rivers may be fault alignments and this is a serious concern in dam engineering. A proper site investigation looking for faults is a must in any dam engineering.

SEISMIC MOMENT

[Strike-slip mechanism is shown]



M_0 = G A δ

G = Shear modulus
A = Area of faulting
δ = Fault Dislocation

Seismic Energy/ Volume = $\frac{1}{2}\frac{P_0^2}{G}(1-\lambda)^2$

$P_0 = G.\delta_0/D$

$\delta = 2.\delta_0(1-\lambda)$

V = 2LDH

Seismic Energy = ¼ M_0 δ/D

log M_0 = 16.1 + 1.5 M_s ------Interplate Earthquakes

log M_0 = 15.6 + 1.5 M_s ------Intraplate Earthquakes
Kanamori & Anderson(1975)

M_0 is in dyne.cm

M_w = Moment Magnitude = log (M_0) / 1.5 – 10.7

Figure 14

Recognition of active faults:

Faults may be classified as
a) active, b) potentially active, c) uncertain activity and d) inactive

Active fault: These show historical or recent surface faulting with associated strong earthquakes. There may be other indications for fault movements such as geomorphic features characteristic of active fault zones along the fault trace.

Potentially active faults: No reliable report of historic surface faulting but geological settings suggest activity similar to nearby active faults.

Faults of uncertain activity: Not enough data available to establish fault activity.
Inactive faults: No activity based on a thorough study. Geological evidence exist to suggest that the fault has not moved in the recent geological past.

Activity of faults may be assessed geologically and seismically.

<u>The Site Parameters</u>:

The effect of the source of the earthquake is transmitted to the site by seismic waves. There are basically two kinds of waves- the **body waves** and the **surface waves**. In an infinite homogeneous medium, only the body waves can be present. Surface waves are generated in the presence of a free surface or along the boundaries of heterogeneous medium.

There are two kinds of body waves:

P waves - These are the compression waves (same as sound waves), propagated by the compression and rarefaction of the medium. The particle motion in these waves is along the direction (ray) of the wave. The velocity of these waves are the highest.

$$V_p = \sqrt{\frac{E(1-\nu)}{\rho(1+\nu)(1-2\nu)}}$$

E= Young's Modulus, ν = Poissons' Ratio of the medium and ρ = mass density.

S waves - These are the shear waves, propagated by the shear action of particles. The particle motion in these waves is perpendicular to the direction (ray) of the waves. The vector of this particle motion can be broken up into two components- one on a vertical plane - called SV component and the other on a horizontal plane - called SH component.
The velocity of these waves is somewhat smaller than the P waves.

$$V_s = \sqrt{\frac{G}{\rho}} = \sqrt{\frac{E}{\rho.2(1+\nu)}}$$; G= μ = Shear Modulus

This gives

$$\frac{V_p}{V_s} = \sqrt{\frac{2(1-\nu)}{(1-2\nu)}}$$ > 1.0 always . For ν=0.25, V_p/V_s = $\sqrt{3}$ = 1.73

There are basically two kinds of surface waves:

Rayleigh Waves : The particle motion in these waves is somewhat similar to the ripples in water (but not exactly the same)- The motion behaves like a combination of P and SV waves, when the direction of the wave is horizontal.
Love waves: The motion behaves like a combination of P and SH waves.

The Rayleigh waves can exist in a homogeneous finite medium. Love waves exist only in heterogeneous medium. The velocities of these waves are smaller than the S waves.

See figure 15.

The reflection and refraction in a boundary of two materials of one kind of incident body waves may generate both kinds of body waves. The partitioning of the incident energy into the four components, two in reflection and two in refraction, depends on the incident angle and on the relative properties of the two media. That is why the earthquake ground motion is very complex.

Because of the difference in velocity of these various waves, different waves arrive at different times at the site. See figure 16. Therefore, knowing the velocity profile of the earth, it is possible to estimate the distance of the source to the site. Seismologists use this information from many sites to locate the epicentre and the focal depth of the earthquake. Since there are four unknowns in the location of epicentres i.e. the latitude, the longitude, the focal depth and the origin time of the earthquake, a minimum of 4 stations is required to locate the source of the earthquake. International Seismological Centre (ISC) in Newbury, Berkshire is equipped to collect the station information from all over the world and determine the hypocentre with as much accuracy as possible by using least-square fitting technique. NEIC (National Earthquake Information Centre- USGS) in the USA is another similar centre. There are centres in every country which collects data from stations in that country and determine epicentres. NEIC also determines magnitudes and moments of earthquakes. ISC usually does not determine magnitudes and moments but reports those given by NEIC and other stations.

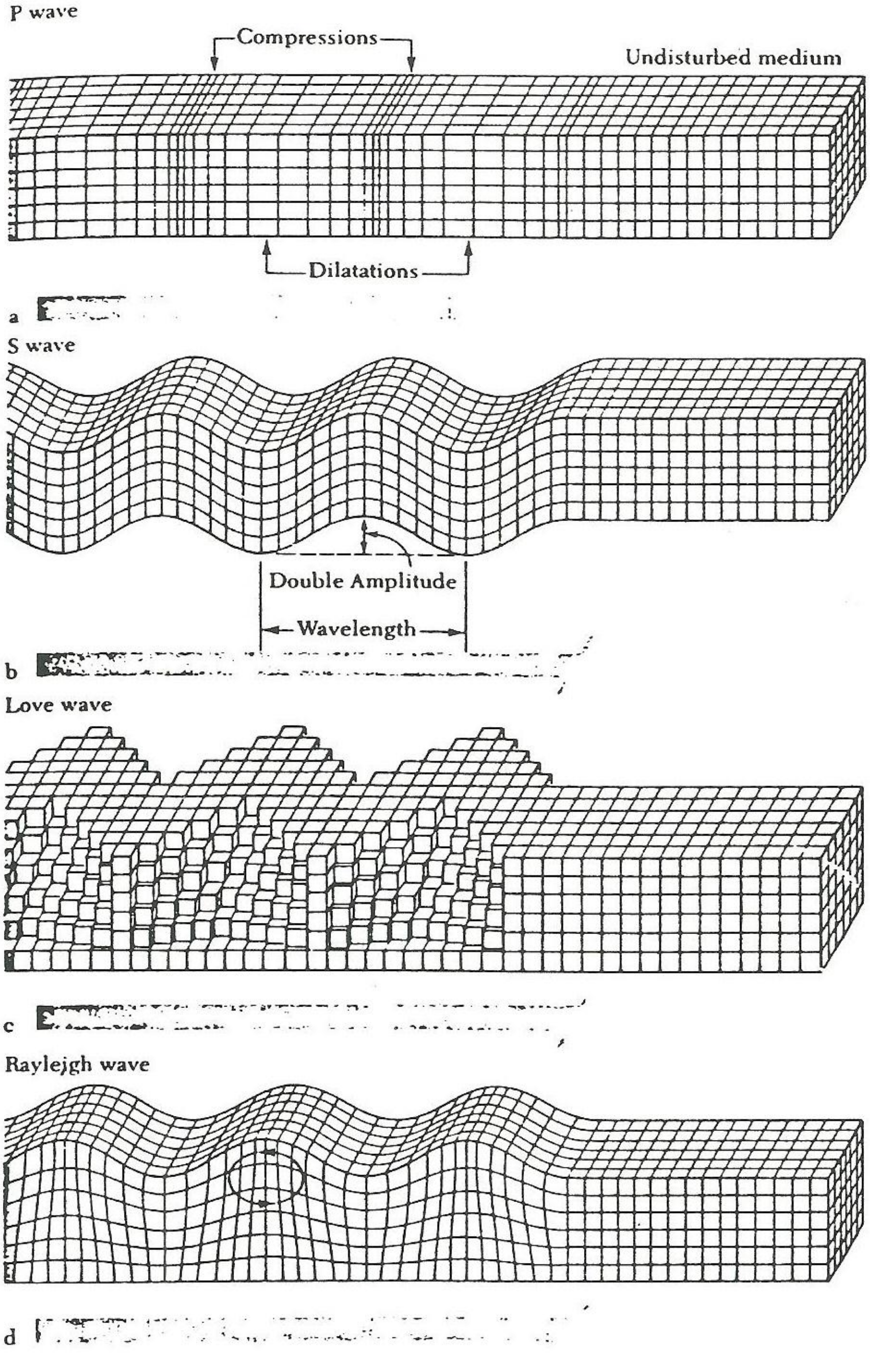


Figure 15: Diagram illustrating the form of ground motion near the ground surface in four types of earthquake waves. [From: Bruce A Bolt, Nuclear explotions and earthquakes, W H Freeman and Company, 1976]

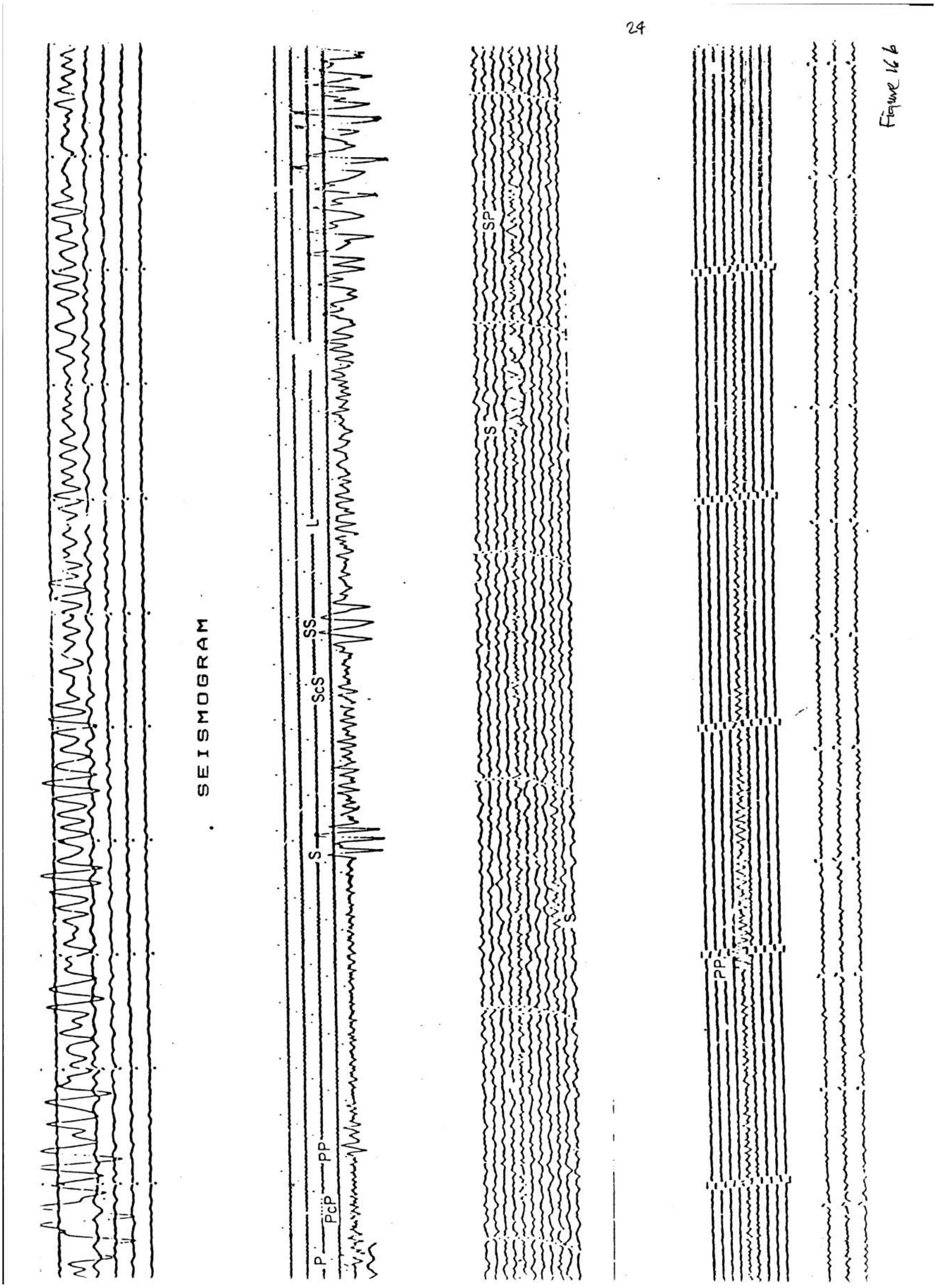


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Figure 16

ELASTIC WAVES

1. BODY WAVES ----- P Waves---Compression waves

------ S Waves---Shear Waves

2. SURFACE WAVES—R Waves--- Rayleigh Waves
--L Waves--- Love Waves

[Reflection and/or Refraction of one kind of body wave at the interface of different materials produces both kind of body waves.]

ATTENUATION OF ELASTIC WAVES

(Decrease of amplitude of waves with distance)

Geometric Attenuation- Due to the expansion of the wave front with distance.

Inelastic energy loss- Due to the irrecoverable work done by particles during vibration.

Figure 17

Seismic Energy at a site

The propagation and attenuation of seismic energy: Figure 18 and Figure 19.
The energy released at the source is propagated by the seismic waves in the form of particle motion. In an infinite medium, the propagation will take place in all direction equally from the source. This is known as the spherical propagation. In this case, the energy of the source is spread around the expanding wave front. In this case, the wave front is the surface area of the sphere that expands with the distance. Therefore, the energy per unit area of the wave front becomes smaller. The site which exists in the wave front will feel this energy. This reduction of energy from source to the site is known as geometric attenuation. The spreading of energy can be in a cylindrical front (for example, if the fault breaks instantly as a line, the spreading will be cylindrical). It can be on a plane front as well in which case there is no geometric attenuation. In reality, geometrical attenuation is a mixture of all kinds.

Besides the geometric attenuation, there is also the energy loss due to the inelastic work done during the particle motion. This is caused by the inter-particle friction but this loss is represented by the viscous damping characteristics (strain rate effect). Due to the viscous damping, the particle motion decreases with distance. The factor by which the ground motion decreases with distance is given by $e^{-\lambda\Omega s/S}$. In this expression, λ is the viscous damping coefficient as a fraction of the critical, Ω is the circular frequency of the wave in radians, s is the distance travelled and S is the wave velocity. In seismology, the value λ is represented by the Q factor (Q=Quality) where $Q = 1/2\lambda$.

The effect of the seismic energy at the site is measured indirectly in two ways:

a) Intensity of earthquakes and b) Ground motion parameters.

Intensity of Earthquakes:

Intensity of earthquakes is a measure of the damage to structures, grounds, slopes etc. and the way human beings and animals react to the earthquake. This is a subjective measure and therefore can be in error, particularly when comparing notes of different observers. When comparing effects on a particular class of structure, the measure could be very effective. But by mixing different class of structures or ground effects, the measures could be confusing. It is even more confusing when slope failures are taken into account. Slopes do fail even without earthquakes. The failure depends on the available factor of safety at the time of the earthquake which depend on many seasonal factors. Therefore, to use the failure of slopes to measure the size of the earthquake is not correct. Intensity serves an important purpose, particularly when assessing pre-instrumental historical earthquakes.

ATTENUATION RELATIONSHIP (IN TERMS OF ENERGY)

$E_s = E_o e^{-kR}/CR^n$

$\log Eo = a + b M$

$\log Es = c + d I$

THUS, $I = B_1 + B_2M + B_3R + B_4 \log R$

WHERE B_2 SHOULD BE A POSITIVE NUMBER AND B_3 AND B_4 SHOULD BE NEGATIVE

ALSO, ENERGY IS PROPORTIONAL TO SQUARE OF GROUND MOTION

THEREFORE, GROUND MOTION y IS OF THE FORM

$y = b_1 e^{b2M} e^{b3R} R^{b4}$

OR $\log y = C_1 + C_2M + C_3R + C_4 \log R$

WHERE C_2 SHOULD BE POSITIVE AND C_3 AND C_4 SHOULD BE NEGATIVE.

GROUND MOTION CAN BE MEASURED IN TERMS OF ACCELERATIONS, VELOCITIES OR DISPALCEMENTS. THE VALUE OF THE CONSTANTS WILL CHANGE WITH THE GROUND MOTION PARAMETER

Figure 18

DEFINITION OF R

THERE ARE MANY ATTENUATION RELATIONSHIPS IN WHICH R IS DEFINED IN MANY DIFFERENT WAYS

EPICENTRAL DISTANCE $R = D_e$

FAULT DISTANCE $R = D_f$

HYPOCENTRAL DISTANCE $R = \sqrt{(D_e^2 + h^2)}$

SLANT DISTANCE TO THE FAULT $R = \sqrt{(D_f^2 + h^2)}$

A BEST FIT h TERM SAME FOR ALL EARTHQUAKES $R = \sqrt{(D_e^2 + ho^2)}$

$R = \sqrt{(D_f^2 + ho^2)}$

A BEST FIT CONSTANT ADDITIVE TERM $R = \bar{R} + C$

WHERE $\bar{R}$ COULD BE ANY OF THE R TERMS ABOVE AND C IS A BEST FIT CONSTANT

A BEST FIT ADDITIVE TERM DEPENDENT ON MAGNITUDE $R = \bar{R} + C_5 e^{C6M}$

Figure 19

There are several intensity scales that are presently in use. The most common is perhaps the Modified Mercalli Scale, developed originally by Mercalli in 1902 and later modified by Wood (1932). The most common scale used in Europe is MSK scale (Medvedev, Sponheur, Karnik). The scales are more or less similar.

After an earthquake, Intensity data is collected and plotted in a map, figure 20. In this map, contours of equal Intensity is drawn which are known as Isoseismals. Generally, isoseismals are not circular, quite often showing signs of high intensity in low intensity regions, mainly due to soil effects. From these isoseismals, an average radius can be computed. From the size of the average radii for different levels of isoseismals, the magnitude of the earthquake can be assessed. See attenuation relationship in terms of Intensity, Figure 21.

The Modified Mercalli Scale:

I Not felt except by a very few under especially favourable circumstances.
II Felt only by a few persons at rest, especially on upper floors of buildings. Delicately suspended objects may swing.
III Felt quite noticeably indoors, especially on upper floors of buildings but many people do not recognise it as an earthquake. Standing motor cars may rock slightly. Vibration like passing trucks. Duration estimated.

IV During the day, felt indoors by many and outdoors by few. At night some awakened. Dishes, windows, doors disturbed; walls make creaking sound. Sensation like heavy truck striking building. Standing motor cars rocked noticeably.
V Felt by nearly everyone; many awakened. Some dishes, windows etc. broken; a few instances of cracked plaster; unstable objects overturned. Disturbances of trees, poles and other tall objects sometimes noticed. Pendulum clocks may stop
VI Felt by all; many frightened and run outdoors. Some heavy furniture moved; a few instances of fallen plasters or damaged chimneys. Damage slight.

VII Everybody run outdoors. Damage negligible in buildings of good design and construction; slight to moderate in well-built ordinary structures; considerable in poorly built or badly designed structures; some chimneys broken. Noticed by persons driving motor cars.
VIII Damage slight in specially designed structures; considerable in ordinary substantial buildings with partial collapse; great in poorly built structures. Panel walls thrown out of frame structures. Fall of chimneys, factory stacks, columns, monuments, walls. Heavy furniture overturned. Sand and mud ejected in small amounts. Changes in well water. Disturbs persons driving motor cars.
IX Damage considerable in especially designed structures; well designed frame structures thrown out of plumb; great in substantial buildings with partial collapse. Buildings shifted off foundations. Ground cracked conspicuously. Underground pipes broken.
X Some well-built wooden structures destroyed; most masonry and frame structures destroyed with foundations; ground badly cracked. Rails bent. Landslides considerable from river banks and steep slopes. Shifted sand and mud. Water splashed over banks.

XI Few, if any, (masonry) structures remain standing. Bridges destroyed. Broad fissures in ground. Underground pipelines completely out of service. Earth slumps and land slips on soft ground. Rails bent greatly.
XII Damage total. Waves seen on ground surfaces. Lines of sight and level distorted. Objects thrown upward into the air.

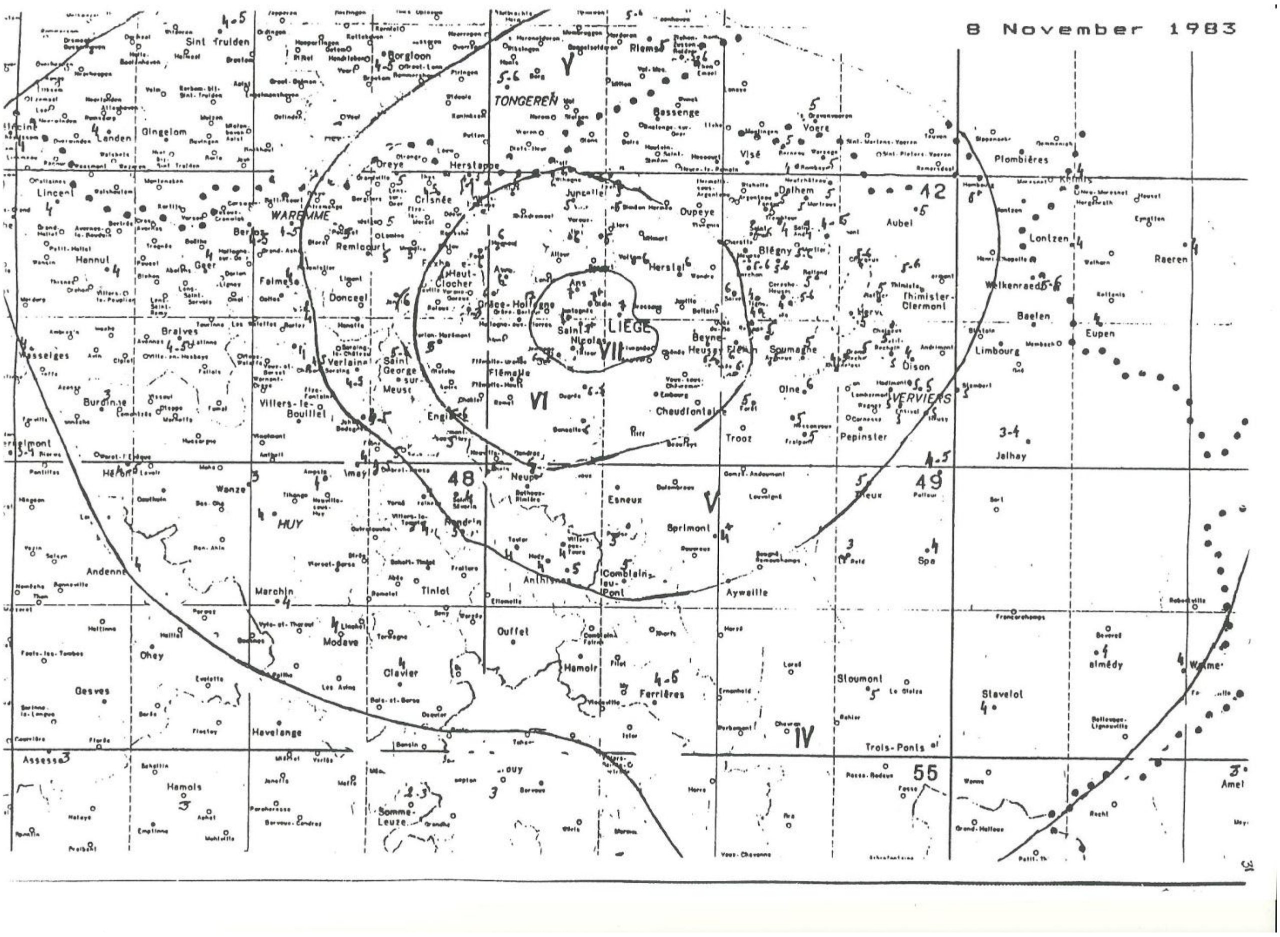
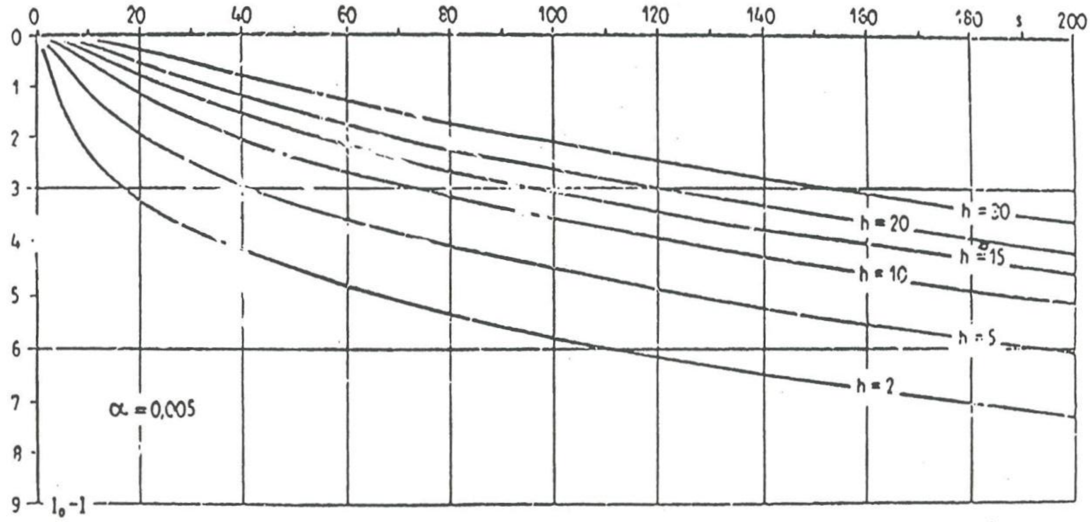


Figure 20



Anlage 10. $I_0 - I$ in Abhängigkeit von s nach (1.6)

Figure 21

Ground Motion:

The effect of the energy at the site can also be measured by the ground motion parameters. The ground motion is measured in terms of accelerations in the near field of earthquakes by using the "Strong motion" instruments. The engineers use these. Seismologists use ground displacements to determine epicentres and other source parameters. These are measured by "Seismographs". Both strong motion instruments and seismographs produce time-history records.

Theoretically, acceleration records can be integrated to obtain ground velocity and displacements. However, there are problems associated with the integration process coming from the "noise" in the records. Therefore, often the displacements obtained from integration process is not reliable. (We do however use them after filtering out the noise, but filtering process is not perfect.) Similarly, the displacement record obtained from seismographs can be differentiated numerically to obtain velocity and accelerations. However, the numerical differentiation process is always inaccurate when spikes are involved in the records. See figure 22.

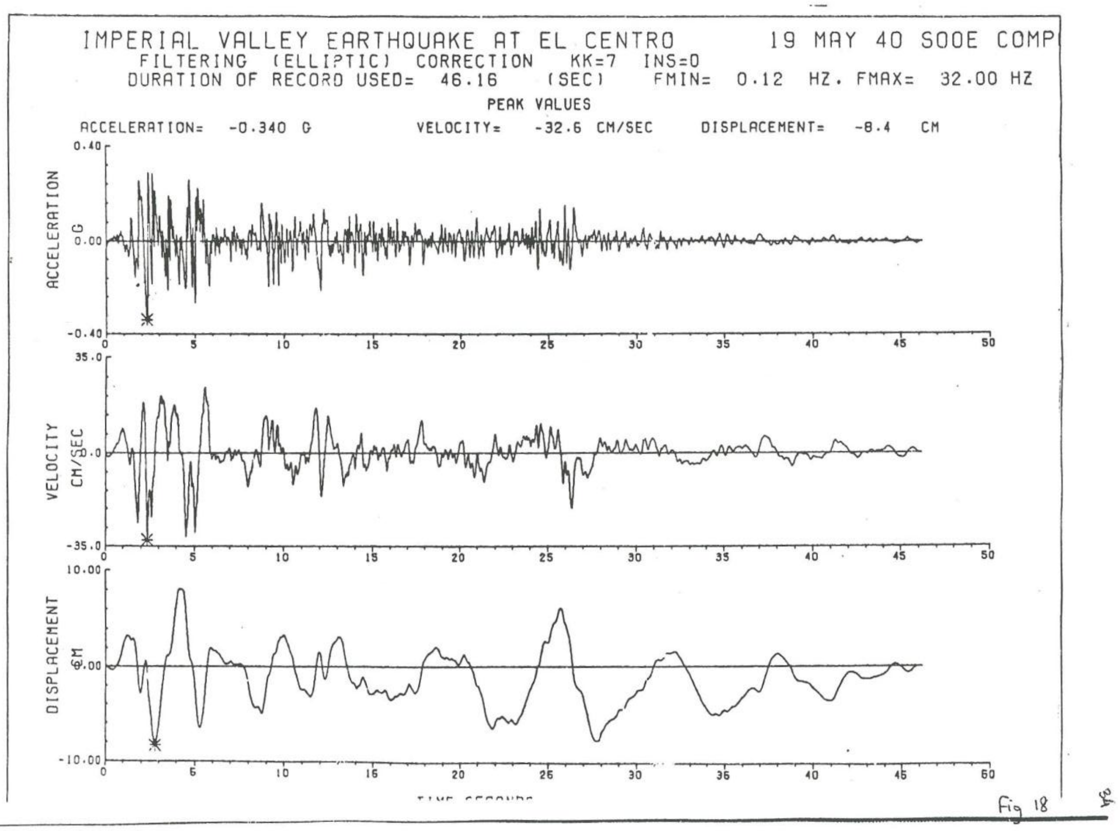


Figure 22

Now-a-days, banks of strong motion records exist (For example,ISESD, the strong motion Data Bank originally stored at Imperial College) where, records from all over the world are collected and processed. Because of their engineering importance, some owners of records tend not to give the records away.

Attenuation of Strong motion Data:

Using the bank of data, attenuation relationships are derived for different ground motion parameters by various authors. These relationships differ because of the choice of data from the bank and the choice of the type of relationship. There are global relationships and there are relationships derived from data from single countries. There are also relationships derived from data of similar tectonic environments or of similar site geology. There can be many such classifications. Because of the complex nature of the strong motion records, the relationships appear to be crude with large standard deviations. Many attempts are going on to obtain better relationships but not with much success. See figure 23.

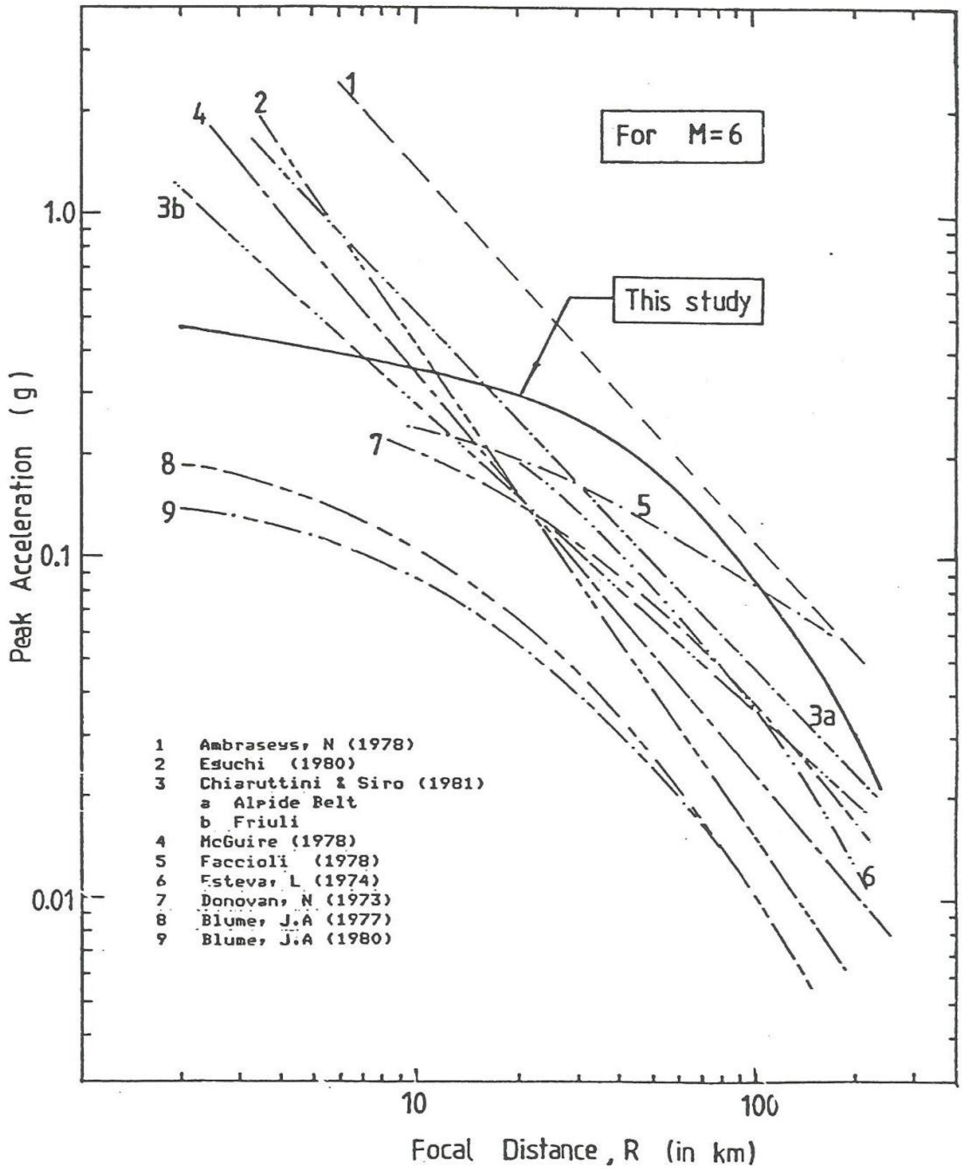


Figure 23

The two most common relationships that are in use are

1. Ambraseys N. and Bommer J.(1991):

$\log a(g) = -1.09 + 0.238\ M_s - \log r - .0005r + 0.28p$; where $r=\sqrt{(D_e^2+ 6^2)}$
p=0 for mean value
p=1 for mean + 1sd
D_e = Shortest distance to the fault rupture.

2. Joyner W.& Boore D. (1981):

$\log a(g) = -1.02 + 0.249\ M - \log r - .00255r + 0.26p$; where $r=\sqrt{(D_e^2+ 7.3^2)}$
Other definitions are same.

Sarma & Srbulov (1998) defined attenuation relationships for other ground motion parameters.

Figure 24 shows that the Intensity of earthquakes and the peak ground acceleration do not really have any correlation even though there is a trend. Any correlation found in the literature between these parameters should be treated with extreme caution.

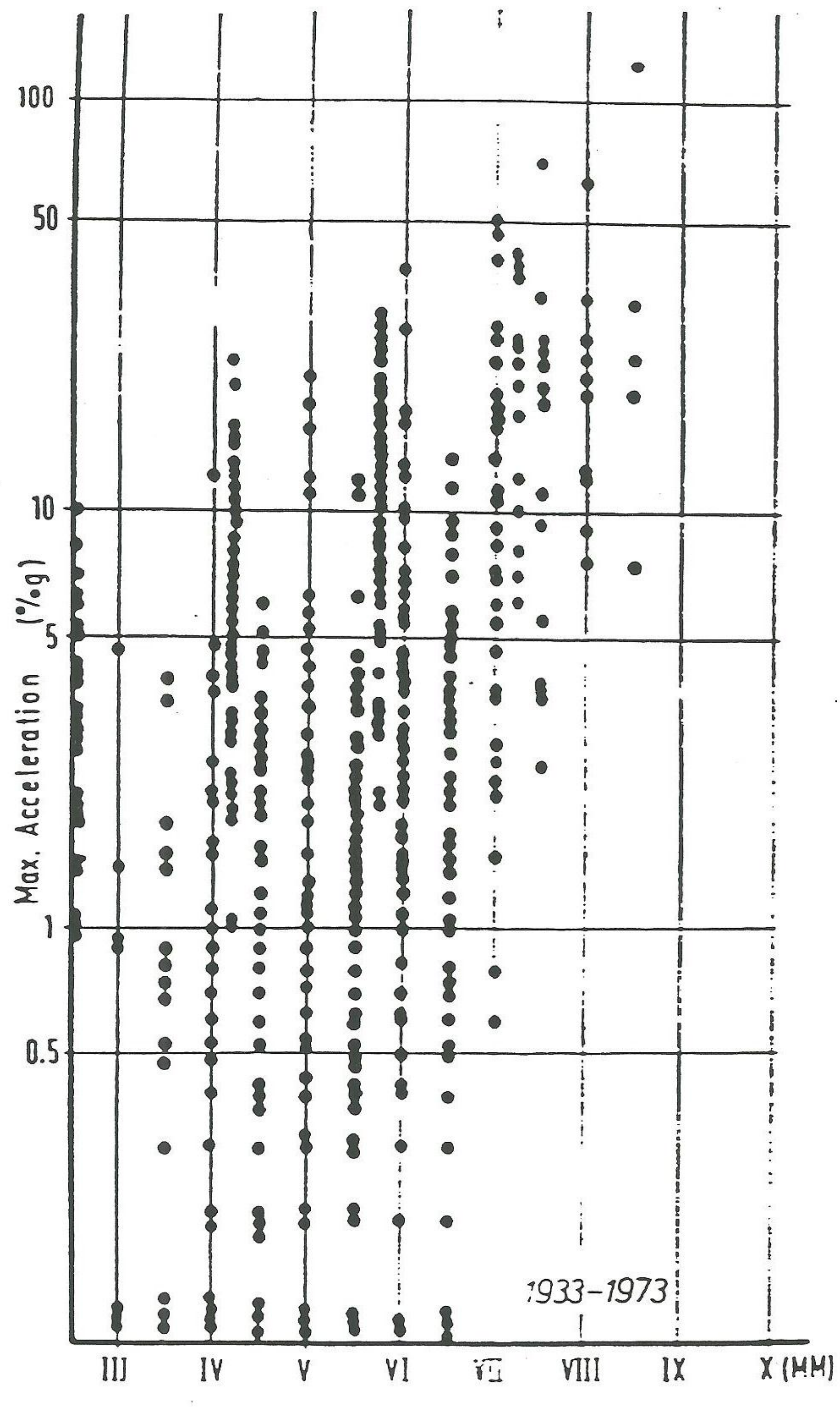

100
50
10
5
1
0.5
Max. Acceleration (%g)
1933–1973
III
IV
V
VI
VII
VIII
IX
X (MM)

Figure 24

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Appendix

SEISMICITY & HAZARD EVALUATION OF A SITE

In order to evaluate the seismic hazard of an engineering site we need the following information.

a) Historical seismicity of the region;
b) Geology and tectonics of the region;
c) A mathematical (statistical) model for analysis;
d) Local soil conditions at the site.

In general, the hazard analysis concerns with the first three factors while the local soil conditions are considered as a special case if necessary.

The study begins with the establishment of the region of interest around the site, which in general could be large, say 5°x5° or even bigger. The idea is to establish regions within this area which can be called homogeneous in the seismic sense, i.e. that the region belongs to the same tectonic province, the earthquakes within the area has the same sort of mechanisms.

Historical seismicity of the region:

For the area in question, we then collect all the data about the earthquakes, i.e. the size (magnitude, moment), location (epicentre, focal depth) that has happened in the past. The data can be divided into two groups, instrumental data and pre-instrumental historical data.

The instrumental data can be obtained from the International Seismological Centre (ISC in UK), the National earthquake Information Centre (NEIC in USA) and the National Geophysical Data Centre(NGDC in USA). These agencies can supply data covering the period from 1906 to the present. The accuracy associated with the instrumental data varies with time. At the early stage, the errors associated could be large, particularly with epicentre determination (±25km). There are instances of gross errors in locations. The reason being very few instruments, unevenly located around the world and with low sensitivity. The location errors in the present time could be ±5km. In the early period, smaller events were not located and therefore incomplete.

The locations of pre-instrumental period earthquakes are obtained from historical studies, extending the period as far back as possible. Obviously, the historical earthquakes will concentrate on the large events. The magnitudes are determined from macro-seismic information, such as felt radius or epicentral Intensity.

The magnitude determination for the instrumental period is non-homogeneous in the sense that different formulae were used in different periods. It is often necessary to recalculate magnitudes in a homogeneous way from the original data or look for published data. The error associated with magnitudes could be of the order of ±0.25.

The focal depth determination is not accurate at all. In the early period, the focal depth was generally given as 'Normal' or 33 km depth. In the catalogues, 33 km depth usually implies

unknown shallow focus earthquake. Even in the present day, errors associated with focal depth determination could be large.
Following the collection of this data, map the epicentre locations, distinguishing between the instrumental and the historical ones and distinguishing the size.

Geological and tectonic data:

We map the known fault location within the area, particularly the active faults, which moved in the quaternary period.

The combination of the two maps will give us an idea of the source region of earthquakes within the area. The sources therefore could appear to be points, lines(faults) and areas. The area source appears due to the uncertainty of the location of faults and the association of faults with epicentres. The source region determination is subjective and not conclusive.

Activity of the source regions:

It has been found that the activity of a source region follows a relationship, Guttenberg & Richter(1954)

$$\log(N_c) = a - bM$$

where M is the earthquake magnitude and N_c is the number of earthquakes of magnitudes greater than or equal to M. In general, the numbers are normalised to a **year** and to **unit area** for area source and **unit length** for linear source. 'a' is therefore a measure of the activity of the region, when normalised. 'b' is a measure of the 'brittleness' of the region. If the crust is highly faulted so that there exists many small faults and few large faults, then 'b' will be large. There will be a tendency for many small earthquakes compared to large earthquakes. The value of b lies between 0.5 to 1.5. Considering the activity of the whole earth, b value is approximately equal to 1.

Due to the incompleteness of the data, deviation from the linear trend exists. We generally do not consider magnitudes less than about 4 in the trend analysis. Also for the highest magnitude, since the period of the catalogue is very limited, this may have to be discarded in the trend analysis.

Maximum magnitude:

For any region, we expect a maximum magnitude. It is essential to assess this maximum magnitude. From the study of the past earthquakes and the tectonic activity of the region, this can be estimated. In the absence of such a study, the largest historical earthquake + a small increment (0.5) is generally considered.

Statistical model:

The statistical model generally applied in hazard analysis is the Poisson process. The Poisson process is memoryless, which implies that earthquakes in one period of time does not depend on the past. This is therefore an assumption. However, it is acceptable for normal hazard analysis. When the hazard is controlled by the very large earthquakes, this assumption may lead to errors.

Return periods:

The return period of an event is simply the average time between events in the past and is given by the inverse of the annual frequency. If n is the number of **favourable events** per year, then the return period of the same is

$$T = 1/n$$

Probability of exceedence:

This is the probability of **at least one favourable event** in the life time of the structure. This is given by the expression

$$p = 1 - \exp(-L/T)$$

where

p is the probability of exceedence
L is the life time of the structure
T is the return period of the favourable event.

Exp(-L/T) represents the probability of non-exceedance.

Attenuation model:

To convert the seismicity information to the ground motion, we need an attenuation model. This model should reflect the geology and the tectonics of the area. For example, the attenuation for Intraplate earthquakes are different from that of Interplate earthquakes. It is preferable to have attenuation relationship for the particular area of concern. This relationship is the most important in the final result and should be chosen with care. Attenuation

$$y = b_1 e^{b_2 M} r^{-b_3} e^{-b_4 r}$$

$$r = \sqrt{d^2 + h^2}$$

relationship for ground motion is of the general form:

There are other forms of r as well such as

$$r = (d + c)$$

b_i are constants dependent on regions.

<u>Hazard Evaluation</u>:

<u>Point Source Model</u>: This model is the basic "building block" for more elaborate source model such as a fault line source or an area source. In this model, a point source with an expected recurrence relationship (a,b parameters) is situated at a given distance (R) from the site and an attenuation relationship exist for the region.

For the point source model, there are two approaches that can be adopted for the analysis.

A) <u>Direct approach</u>: Given the expected life (L) of a structure and the acceptable probability of exceedance (p), we can determine the return period (T) of the event. Thus

$$p = 1- \exp(-L/T)$$

The return period (T) is the inverse of the average number (n) of earthquakes per year.

$$T = 1/n$$

(n) is related to the magnitude of the earthquake through the recurrence relationship

$$\log(n)= a-bM$$

(Note: If the computed magnitude is bigger than the maximum magnitude, then M is the maximum magnitude)

From the magnitude of the event and the distance, we find the design ground motion.

$$y = b_1 e^{b_2 M} R^{-b_3} e^{-b_4 R}$$

(Note: In this relatonship, R is a distance parameter and not the distance directly).

Because of the presence of the maximum magnitude, this approach is applicable for a point source only.

B) <u>Indirect approach</u>: This approach can be extended to more elaborate source models. This is a reverse procedure from the direct approach.

We start with an assumed value of the ground motion y and determine its return period T which is then related to p.

$$y \Rightarrow M \Rightarrow n \Rightarrow T$$

Plot y versus T and determine the design ground motion from the plot.

<u>**Many point sources model**</u>:

In this model, for any given value of the ground motion (y), the (n) values from all point sources are added together. The return period is then given by:

$$T = 1/\Sigma n$$