Earthquakes Mechanics and Effects

This topic explains:

1. The physical mechanisms of an earthquake, plate tectonics, the various types of faults, the earthquake source, path and site elements, and the terminology used to describe the location, severity, and frequency of an earthquake and maps of its physical effects in earthquake-prone areas of the nation and
2. The use of geology, seismicity, and paleoseismicity to determine magnitude as a measure of the “size” of an earthquake and the intensity as the damage state of buildings from ground shaking and ground failure.

References are:

Bolt, B. 1999. *Earthquakes*, 4th Ed. New York, New York: W. H. Freeman and Company.

Gere, J. M., and H. C. Shaw, H. C. 1984. *Terra Non Firma*. New York, New York: W. H. Freeman and Company.

Naiem, F., and J. C. Anderson. 2002. “Probabilistic Methods in Earthquake Engineering.” In *Mechanics for a New Millenium*, proceedings of the 20th International Congress of Theoretical and Applied Mechanics Chicago, Illinois, USA 27 August – 2 September 2000.

Reiter, L. 1990. *Earthquake Hazard Analysis*. New York, New York: Columbia University Press.

Seed H. B. and I. M. Idriss. 1971. “Simplified Procedure for Evaluating Soil Liquefaction Potential, *J. Soil Mech. & Foundations Div.*, ASCE, 97(9), 1249-1273.

Earthquakes: Cause and Effect

- Why earthquakes occur
- How earthquakes are measured
- Earthquake effects
- Mitigation strategy
- Earthquake time histories

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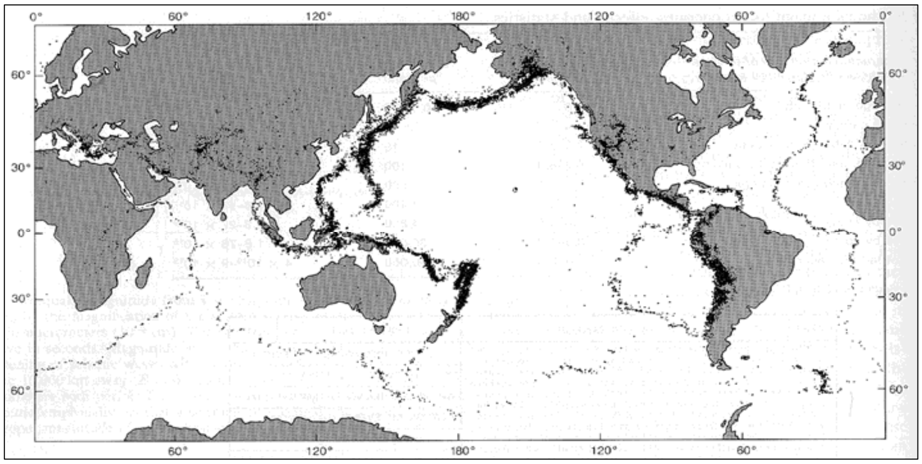
This series of slides introduces the source and effect of ground motions. The information presented is a prelude to the information on the dynamics of single-degree-of-freedom systems, which discusses response spectra, and on seismic hazard analysis.

Seismic Activity: 1961-1967



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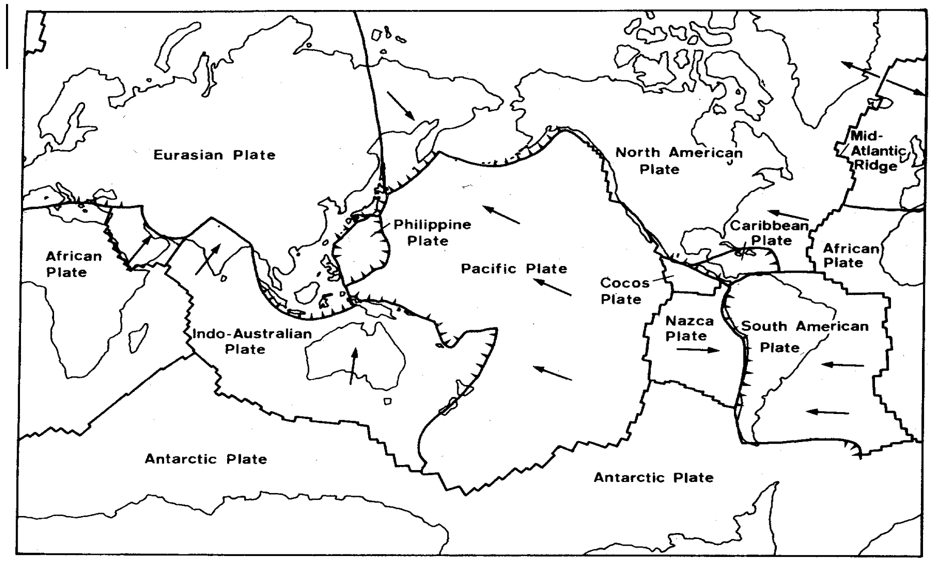
On this map of the world, a small dot drawn at the location of each earthquake that occurred in the years 1961-1967. Note how the dots seem to cluster around the rim of the Pacific Ocean, the mid-Atlantic, and a few other locations. The earthquakes represented by these dots are occurring at so-called plate boundaries -- zones of weakness in the Earth's crust or *lithosphere*. Each of the zones surrounded by the well defined boundaries are referred to as a tectonic plate and the creation, movement, and interaction of these plates is called p*late tectonics*. For lists of recent earthquakes, see the following U.S. Geological Survey (USGS) website: http://earthquake.usgs.gov/eqcenter/index.php.

Plate Boundaries



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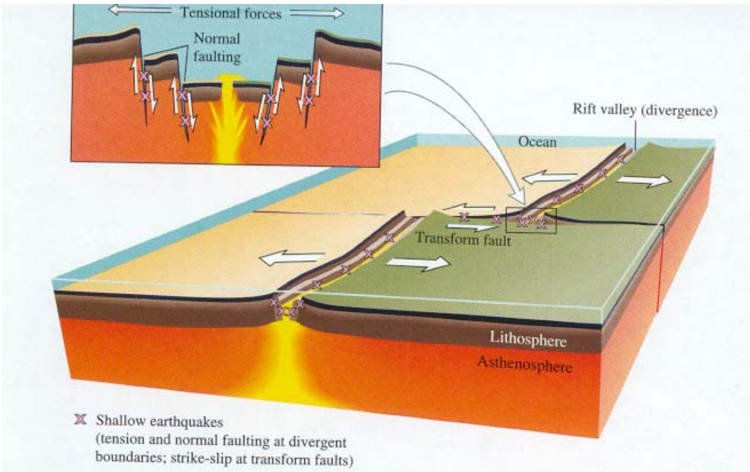
This slide is similar to the previous one except that the plate boundaries are drawn in, the individual plates are named, and the general direction of the movement of the plates is shown. The plates are moving at the rate of 2 to 18 cm (1 to 7 inches) per year with the largest movement occurring in the Pacific Plate. The driving mechanism for the plates is the rising and eventual cooling of molten mantle at mid-ocean ridges and the resulting spreading. The plates do not grow in size because, at some boundaries, the plates are forced to subduct under the thicker continental places, turning the cold rock back into molten mantle.

Plate Tectonics: Driving Mechanism



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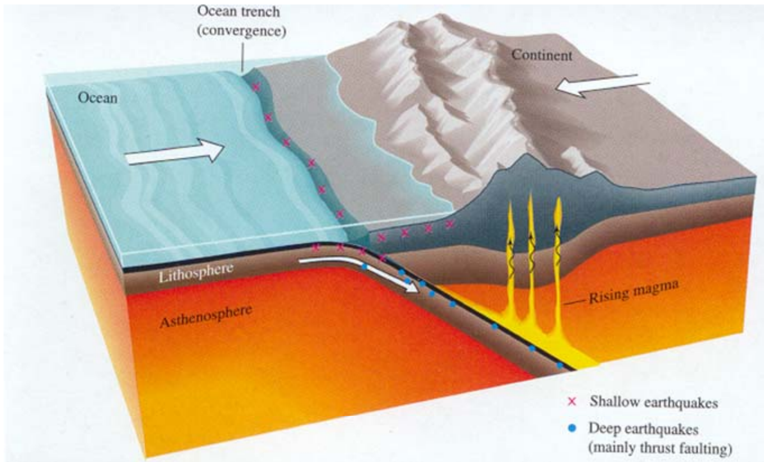
This illustration from Bolt (1999) shows the driving mechanism for plate movement. At this mid-ocean ridge, the mantle rises and cools causing the rift valley divergence and plate separation.

Plate Tectonics: Details in Subduction Zone



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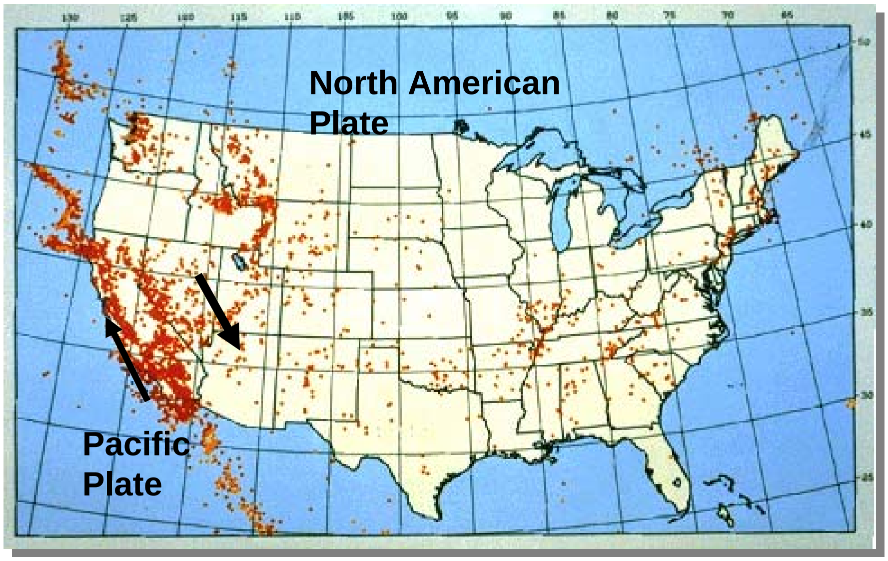
If the plates continue to diverge at the ridges, the surface of Earth would have to grow (or buckle) unless there were some mechanism for returning cool rock into the *asthenosphere*. This slide, also from Bolt (1999), shows how the plate submerges (or subducts) under the continental shelf at the plate boundary. The sudden release of the frictional forces that develop at this interface is a major source of earthquakes. Volcanic activity is also a source of earthquakes but the resulting ground motions are usually not as severe.

Seismicity of North America



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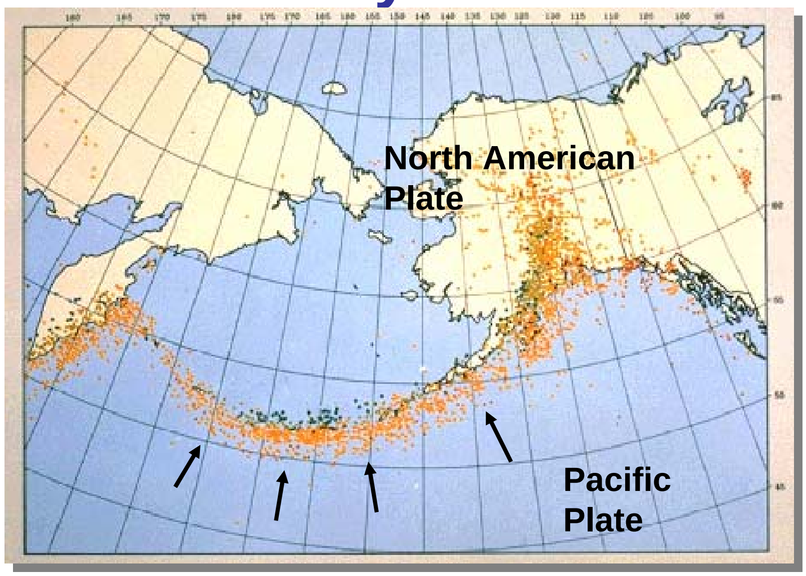
This is a close-up view of the relative movement between the Pacific Plate and the North American Plate. At the boundary of these plates, the two plates move laterally with respect to each other causing very large shear stresses to develop. Over a period of time (10 to 100 years), the stresses in the rock build up until the material ruptures (slips) along zones of weakness or faults. One of the principal zones of weakness at this boundary is the San Andreas Fault, which is the source mechanism for many earthquakes including the 1906 San Francisco earthquake. The type of boundary shown here is referred to as a *transform* boundary.

Seismicity of Alaska



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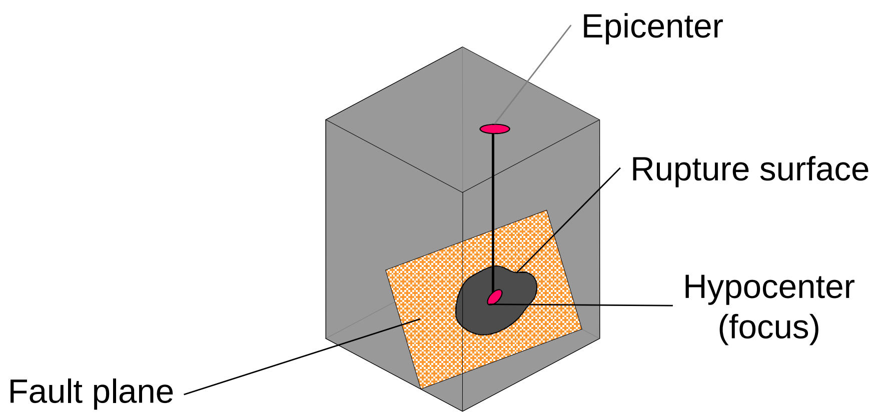
This slide shows how the Pacific Plate and the North American Plate are colliding head-on. The Pacific Plate subducts under the North American Plate creating a *subduction zone*. In subduction zones, the earthquakes are deeper and often much more severe than the earthquakes that occur at San Andreas type boundaries. Aside for the zone shown, other well known subduction zones occur at the boundary of the Nazca Plate and the South American Plate and at the convergence of the Juan de Fuca and the North American Plates in the Pacific Northwest. Subduction zones have generated many of the truly great earthquakes such as those that occurred in 1960 in Chile and in 1964 in Alaska.

Faults and Fault Rupture



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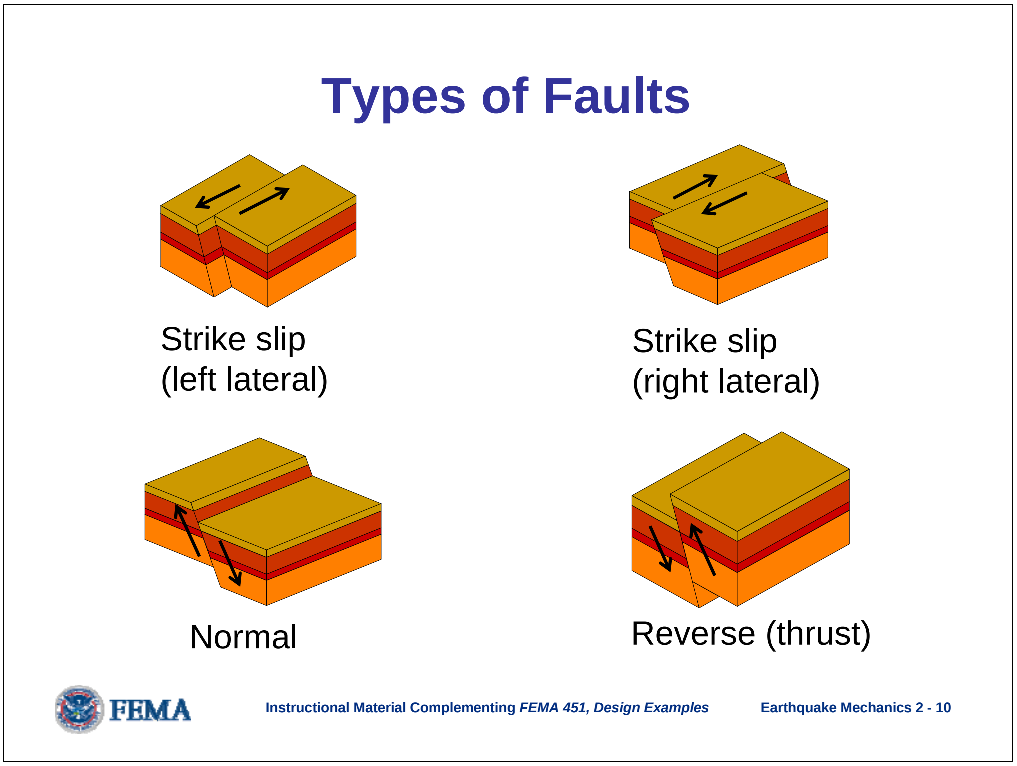
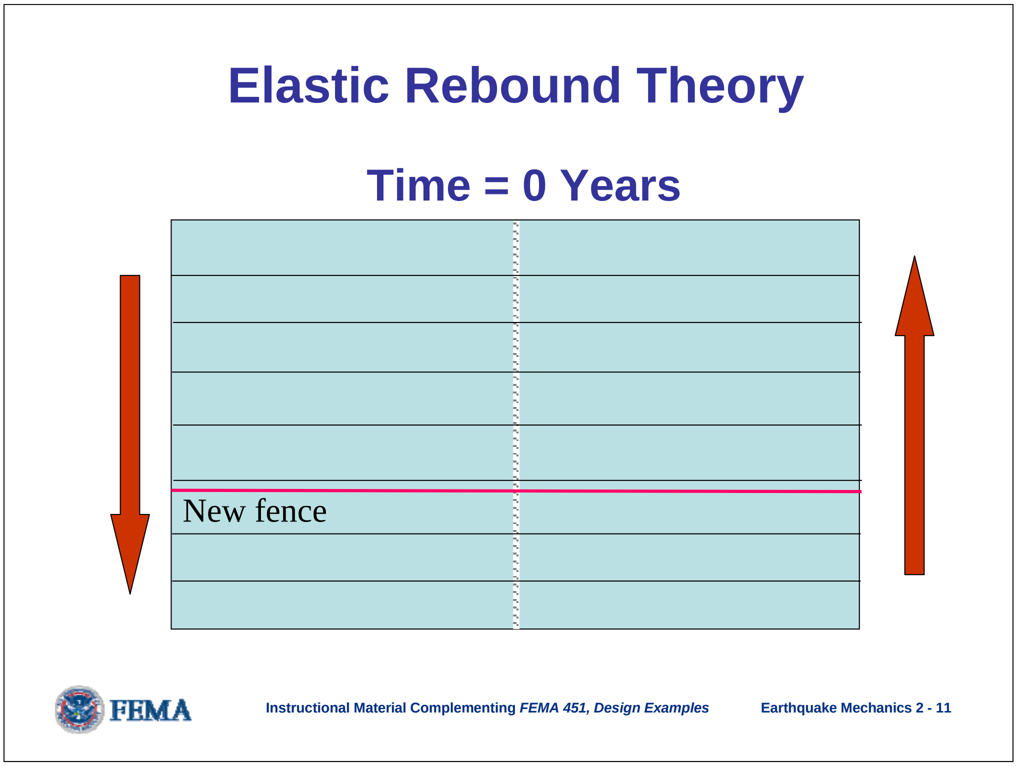
Zones of relative weakness in the Earth's crust are called *faults*. After the stresses build up in the rock, one particular area -- the *focus* or *hypocenter* -- of the fault ruptures first. The rupture then propagates at a very high speed forming a *fault plane*. The vertical projection of the focus to the surface is the *epicenter*. For shallow earthquakes, the fault plate may intersect the surface causing a visible *fault rupture* and possible *escarpment*. For some deeper earthquakes, the fault may not be seen at the surface. These are called *blind faults*.

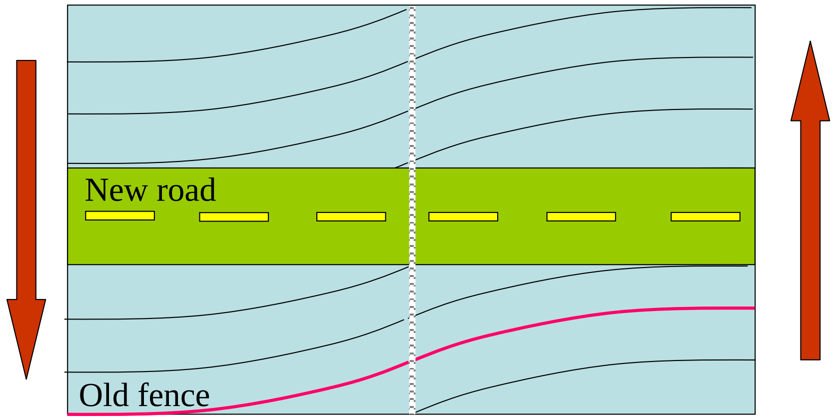


These are the various types of fault representing either lateral or vertical movement. The fault plane may be vertical or skewed as shown. For *strike slip* faults, the type designation comes from the movement of one block relative to an observer. If the observer is standing on one of the blocks looking across the fault and the far block moves to the observer's right, it is a *right lateral* fault. For a *normal fault*, the two blocks move away from each other (extensional). For a *thrust fault*, the blocks moves towards each other (compressional). The visible wall formed from the movement is called an *escarpment*. Of course, the fault may be a combined strike-slip or normal/reverse fault.



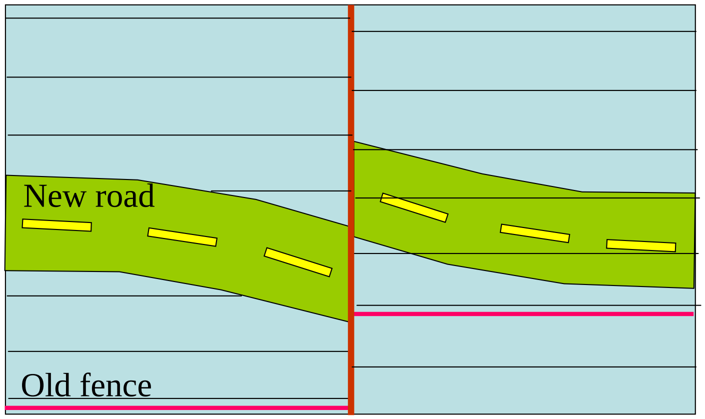
This slide is the first of a series of three illustrating the formation of a left lateral strike slip fault and the elastic rebound that occurs after the fault suddenly ruptures. The observer has a bird's eye view of the a farmer's field. The fault (as yet not ruptured) is the vertical line. The time (shown as zero years) may actually be one to ten years since the last major movement of the fault. The farmer has just built a fence perpendicular to the fault. The thin black lines are for reference only.

Time = 40 Years



After a period of time (40 years) the fault plane itself is still intact, but elastic stresses have built up considerably causing the rock to deform elastically as shown. Note the new shape of the fence as well as the reference lines. The road was recently built and has not yet deformed.

Time = 41 Years



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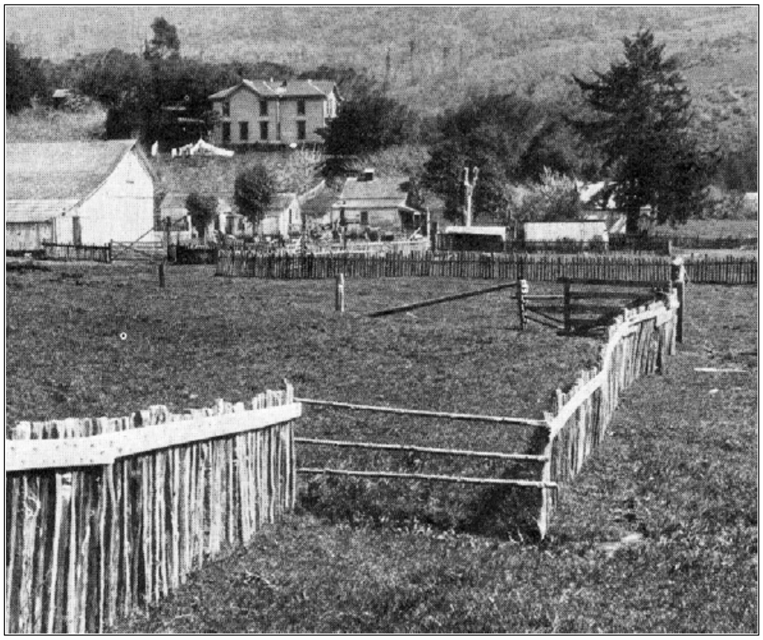
When the strained rock is stressed to its limit, the fault ruptures suddenly releasing a tremendous amount of strain energy. (Recall the noise made by the rupture of a steel coupon or concrete cylinder in the testing lab.) The release of strain energy is accompanied by a variety of seismic waves that propagate in all directions. Note that after the earthquake, the fence is straight but has a significant offset. This offset can be several meters after a major earthquake. The road, which was built straight on deformed land, is now deformed. After a time, the process will begin anew and the stresses will begin building up again.



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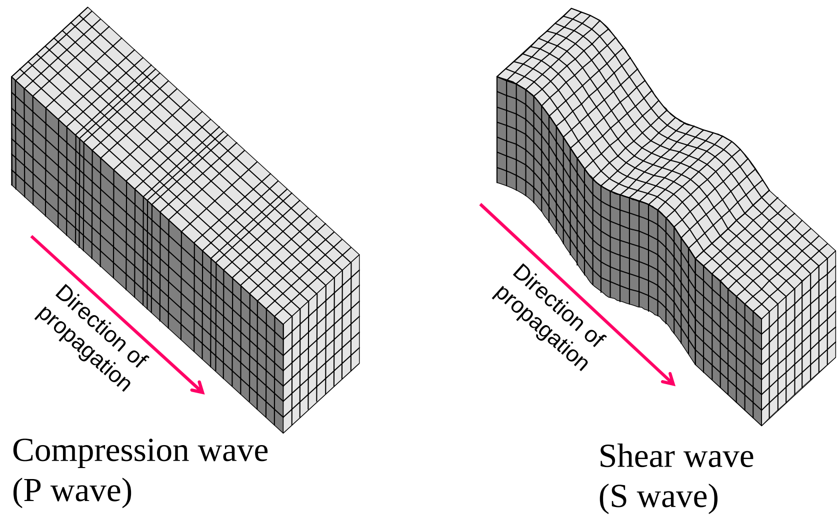
This is a photo of a rupture in the San Andreas Fault after the 1906 San Francisco earthquake. As can be seen, the fault moved approximately eight to ten feet. Interestingly, there is little or no apparent damage to the barn and house shown in the background.

Seismic Wave Forms (Body Waves)



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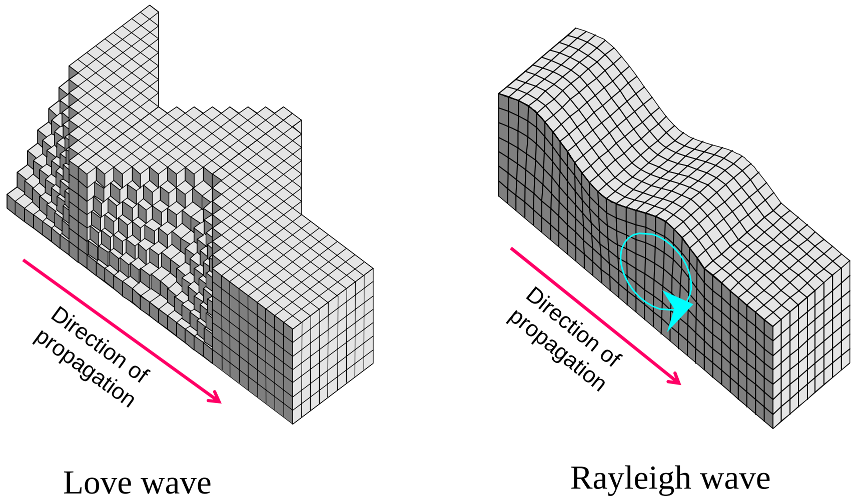
The energy released during an earthquake propagates in waves. The two types of wave are *body waves* and *surface waves*. The principal body waves are the *compression (P) wave* and the *shear (S) wave*. Compression and shear waves move on a spherical front. Sometimes the compression waves are called "push-pull" as they work like an accordion. Compression waves travel the fastest of all waves (4.8 km/second in granite), and they travel through both solids and liquids. Shear waves move from side to side. Because fluids (e.g., water and magma) have no shear stiffness, shear waves do not pass through them. Shear waves are the second wave type to arrive, moving at about 3.0 km/second.

Seismic Wave Forms (Surface Waves)



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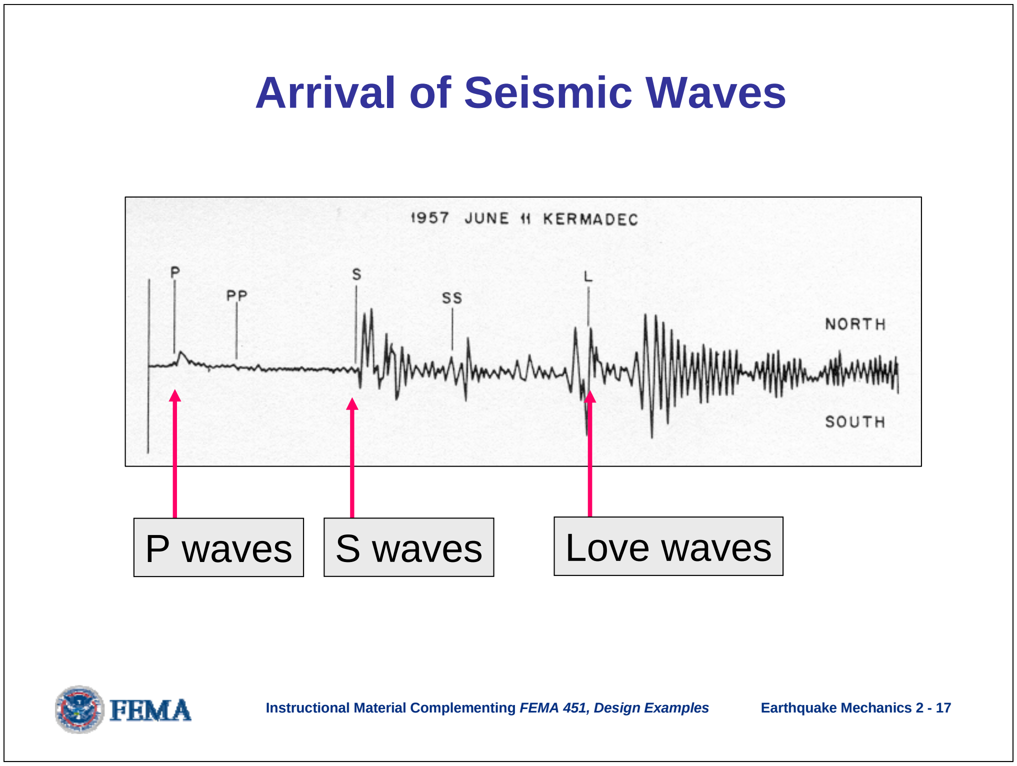
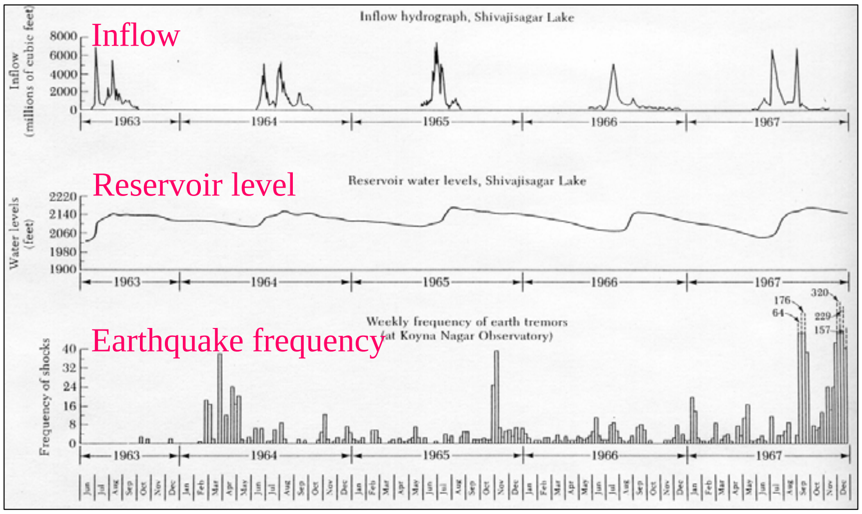
The next waves to arrive are the surface waves. The two main types are *Love* waves and *Rayleigh* waves. These waves have a somewhat longer period than P or S waves.



This recording shows the sequential arrival of P, S, and Love waves. With travel speeds of the various waves known, this type of diagram can be used to estimate the distance to the wave source.

Relationship Between Reservoir Level and Seismic Activity at Koyna Dam, India



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Although the vast majority of earthquakes are due to tectonic effects, there are other causes. One of the more interesting causes is the filling of new reservoirs after a dam has been constructed. The cause of the earthquake is not the weight of the water in the reservoir (which is negligible) but rather the lubricating effect of the water as it permeates previously inactive faults.

This slide shows how the frequency of earthquakes is related to the filling of the Koyna reservoir. Of greatest interest is the large number of tremors in 1967 (hundreds of events per week) compared to the period before filling began (early 1963 when there were zero events per week). Note, however, that it is unlikely that reservoir-triggered earthquakes could cause widespread damage.

Effects of Seismic Waves

- Fault rupture
- Ground shaking
- Landslides
- Liquefaction
- Tsunamis
- Seiches

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Numerous other hazards are related to earthquakes. While ground shaking is emphasized in this topic, it is not necessarily the greatest hazard.

Surface Fault Rupture, 1971 Earthquake in San Fernando, California

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This slide shows the effect of (what appears to be) a surface fault rupture during the 1971 San Fernando Valley earthquake. Clearly, any structure built over such an area would have little possibility of survival. Note that lateral spreading associated with liquefaction can produce similar effects.

Cause of Liquefaction

"If a saturated sand is subjected to ground vibrations, it tends to compact and decrease in volume.

If drainage is unable to occur, the tendency to decrease in volume results in an increase in pore pressure.

If the pore water pressure builds up to the point at which it is equal to the overburden pressure, the effective stress becomes zero, the sand loses its strength completely, and liquefaction occurs."

Seed and Idriss (1971)

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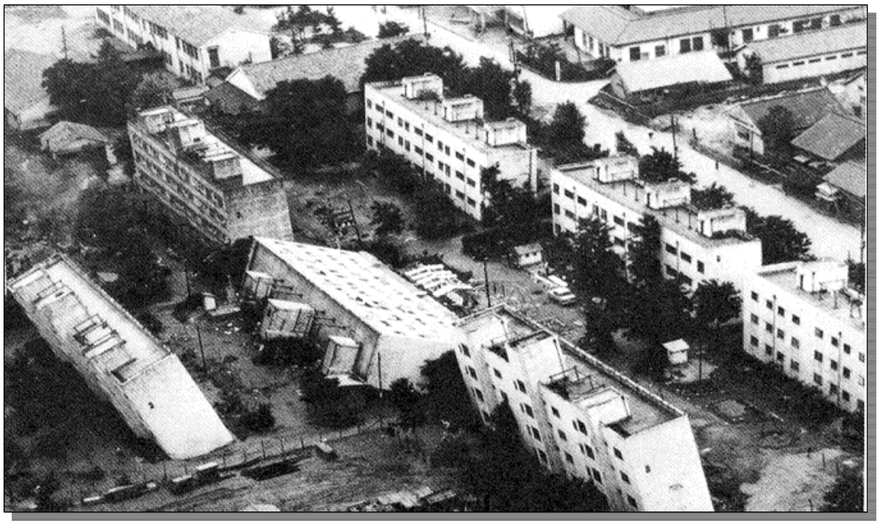
Liquefaction is an extreme hazard for all structures. Liquefaction occurs when saturated cohesionless soils (e.g., sands) are shaken, causing them to compact and decrease in volume. If drainage is unable to occur, the tendency to decrease in volume results in an increase in pore pressure. If the pore pressure builds up to the point where it equals the overburden pressure, the effective stress becomes zero, the sand loses its strength completely, and liquefaction occurs.

Liquefaction Damage, Niigata, Japan, 1964



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This slide shows extreme liquefaction damage. A related phenomena is lateral spreading (due to volume reduction). Liquefaction also can occur in a dam's foundation, at the abutments, or in slopes above and below the reservoir. The Lower Van Norman Dam experienced liquefaction problems during the 1971 San Fernando Valley earthquake.

Effects of liquefaction are often used in the field of paleoseismology. For example, evidence of liquefaction from the New Madrid earthquakes has allowed seismologists to predict recurrence rates (most recent estimate is major event every 300 years). Sand boils form the New Madrid quake are still visible from the air (almost 200 years after the earthquake.)

Liquefaction and Lateral Spreading, 1993 Earthquake in Kobe, Japan

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The decease in volume associated with liquefaction leads to the lateral spreading phenomena shown here. These effects are sometimes mistaken for fault rupture.

Landslide on Coastal Bluff, 1989 Earthquake in Loma Prieta, California

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Landslides are often the result of severe ground shaking. Sometimes the effect is a direct loss of structures and sometimes the loss is indirect, such as flooding caused by landslide materials damming streams and rivers.

Cause of Tsunamis

Tsunamis are created by a sudden vertical movement of the sea floor.

These movements usually occur in subduction zones.

Tsunamis move at great speeds, often 600 to 800 km/hr.

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The December 2004 Indian Ocean earthquake, known by the scientific community as the Sumatra-Andaman earthquake, reminded us that the tsunami's spawned by earthquakes can cause a great number of casualties – in this case, casualties are estimated at 230,000 -- and major property losses. Note that while tsunamis and earthquakes cannot yet be predicted, early warning systems do exist for tsunamis. For example, a tsunami generated by an earthquake occurring 2000 miles away from a shoreline will take approximately four hours to reach that shoreline. This gives emergency preparedness teams some time to warn vulnerable regions. Use of these systems, however, is far from universal.

Tsunami Damage, Seward, Alaska, 1964

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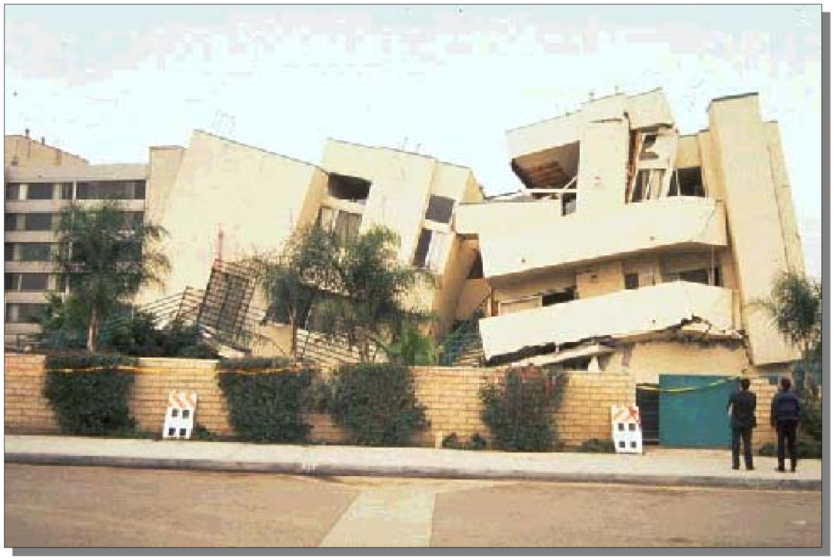
Damage from the 1964 Alaska earthquake. Recall that Alaska quakes are generally subduction earthquakes, and it is this type of ground motion that generates tsunamis.

Result of Ground Shaking, 1994 Earthquake in Northridge, California



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The most well recognized effect of earthquakes is ground shaking, and most damage to property and loss of life are directly related to these effects. The type and extent of damage shown here for a relatively modern building in the Northridge, California, area has caused engineers to re-think the philosophy of seismic design, which is stimulating the current development of performance-based design principles.

Mitigation Strategies

Earthquake effect	Strategy
Fault rupture	Avoid
Tsunami/seiche	Avoid
Landslide	Avoid
Liquefaction	Avoid/resist
Ground shaking	**Resist**

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This slide lists recommended mitigation strategies. These are applicable to dams, bridges, and buildings. The two principal hazards for buildings are liquefaction and ground shaking. Liquefaction may be resisted by utilizing special techniques but, more often than not, avoidance is the best approach. Generally speaking, ground shaking cannot be avoided and must, therefore, be resisted. The *NEHRP Recommended Provisions* deals exclusively with the effects of ground shaking.

Note that fire, which is often associated with earthquakes, is an indirect hazard.

Measuring Earthquakes

INTENSITY
- Subjective
- Used where instruments are not available
- Very useful in historical seismicity

MAGNITUDE
- Measured with seismometers
- Direct measure of energy released
- Possible confusion due to different measures

There are fundamentally two different ways to measure earthquakes. The first, referred to as *intensity*, is a subjective measure that uses damage descriptions from and anecdotal accounts of an earthquake. This measure is often used when instrument data (the second approach) are not available and for earthquakes that occurred prior to the advent of instruments. Intensity is sometimes referred to as historic seismicity. Because of its subjective nature (and different building practices), it is less reliable that instrument records. When instruments are used, they are indirectly measuring the energy released by an earthquake. This energy, in terms of *magnitude*, is called instrumental seismicity. Seismic hazard analysis usually depends on both types of earthquake measure.

Modified Mercalli Intensity

- Developed by G. Mercalli in 1902 (after a previous version of M. S. De Rossi in the 1880s)
- Subjective measure of human reaction and damage
- Modified by Wood and Neuman to fit California construction conditions
- Intensity range I (lowest) to XII (most severe)

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The next series of slides deals with measuring the effects of earthquakes. Both intensity and magnitude will be addressed. The *modified Mercalli scale* is the most common measure of earthquake *intensity*.

Modified Mercalli Intensity

I. Not felt except by a few under especially favorable circumstances.

II. Felt only by a few persons at rest, especially on upper floors if buildings. Suspended objects may swing.

III. Felt quite noticeably indoors, especially on upper floors of buildings. Standing automobiles may rock slightly. Vibration like passing truck.

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The above are the lowest Mercalli intensity numbers. An Intensity III earthquake is very small and would not be felt by many people.

Modified Mercalli Intensity

IV. During the day, felt indoors by many, outdoors by few. At night, some awakened. Dishes, windows, doors disturbed; walls make creaking sound. Sensation like heavy truck striking building. Standing automobiles rocked noticeably. [0.015 to 0.02g]

V. Felt by nearly everyone, many awakened. Some dishes and windows broken. Cracked plaster. Unstable objects overturned. Disturbance of trees, poles and other tall objects. [0.03 to 0.04g]

VI. Felt by all. Many frightened and run outdoors. Some heavy furniture moved. Fallen plaster and damaged chimneys. Damage slight. [0.06 to 0.07g]

These are intermediate Mercalli intensities. Note that at Intensity IV, the scale is indicating a range of peak ground accelerations that can be expected to occur during the earthquake. The accelerations shown on this slide are certainly perceptible to humans.

Modified Mercalli Intensity

VII. Everybody runs outdoors. Damage negligible in buildings of good design and construction, slight to moderate in well built ordinary structures, considerable in poorly built or badly designed structures. Noticed by persons driving cars. [0.10 to 0. 15g]

VIII. Damage slight in specially designed structures, considerable in ordinary construction, great in poorly built structures. Fall of chimneys, stacks, monuments. Sand and mud ejected is small amounts. Changes in well water. Persons driving cars disturbed. [0.25 to 0.30g]

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Intensity VII and VII can produce significant damage as well as injury and loss of life (particularly from falling matter) during an earthquake. However, earthquakes of this intensity do not generally lead to complete system collapses and large-scale loss of life.

Modified Mercalli Intensity

IX. Damage considerable in specially designed structures, well designed frame structures thrown out of plumb, damage great in substantial buildings with partial collapse. Buildings shifted off foundations. Ground cracked conspicuously. Underground pipes broken. [0.50 to 0.55g]

X. Some well built wooden structures destroyed. Most masonry and frame structures destroyed with foundations badly cracked. Rails bent. Landslides considerable from river banks and steep slopes. Shifted sand and mud. Water splashed over banks. [More than 0.60g]

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These are high (but not the highest) intensity numbers and are indicative of a major earthquake.

Modified Mercalli Intensity

XI. Few, if any, (masonry) structures left standing. Bridges destroyed. Broad fissures in ground. Underground pipelines completely out of service. Earth slumps and land slips in soft ground. Rails bent greatly.

XII. Damage total. Waves seen on ground surface. Lines of sight and level distorted. Objects thrown into air.

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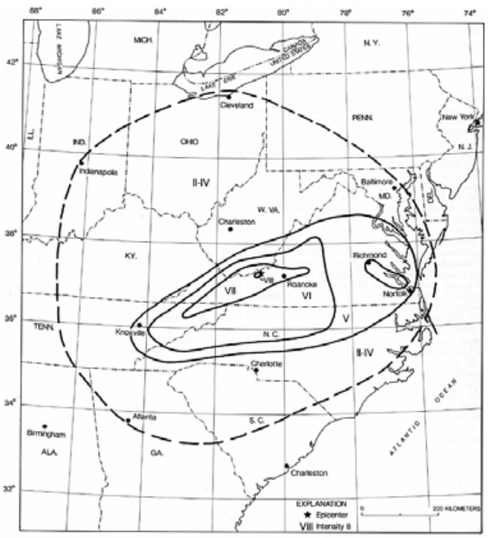
These are the largest intensities and are characteristic of great earthquakes.

Isoseismal Map for the Giles County, Virginia, Earthquake of May 31, 1897.



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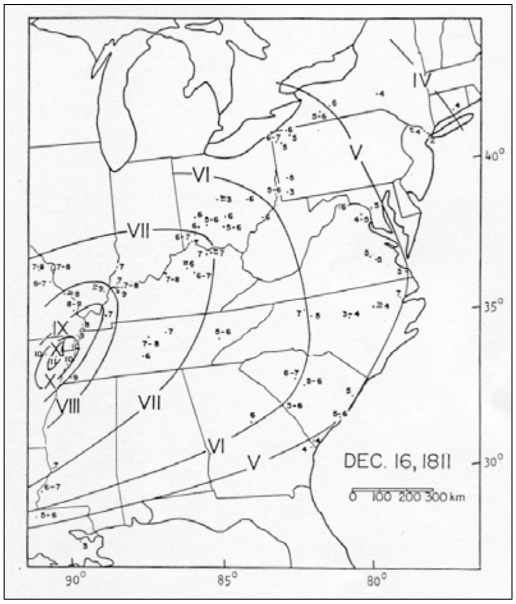
After an earthquake, descriptions of damage and human reaction to the earthquake are recorded over a large geographic area. Using the information, intensities are estimated and then plotted to form isoseismal maps such as the one shown here for Giles County, Virginia. The dotted line on this particular map shows the limit of the "felt area." The intensity at the epicenter (if known) is referred to as the epicentral intensity (typically referred to as *Io*.) Note that because of site amplification effects (described later), the greatest intensity is not necessarily at the epicenter.

Isoseismal maps can be produced immediately after an earthquake or may be produced years later from news accounts, personal dairies and letters, and other historical sources of information. Use of this kind of information is very helpful in the field of seismic hazard analysis when instrument-based measurements (described later) are not available.



Isoseismal Map For New Madrid Earthquake of December 16, 1811

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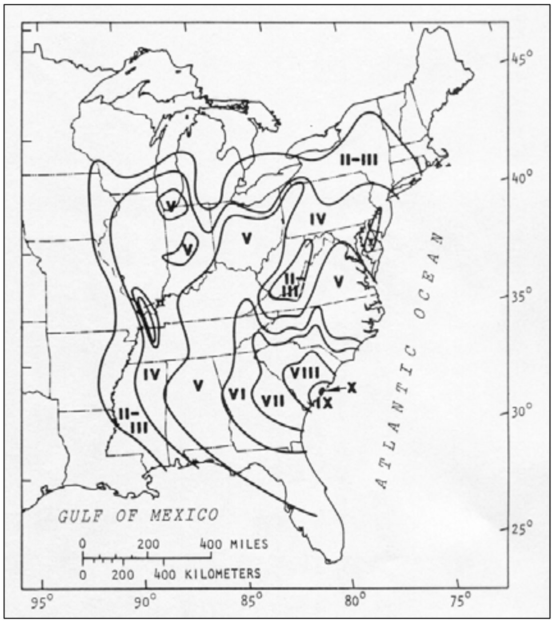
This is an isoseismal map for the New Madrid earthquake. The highest intensity is approximateily XI, which is characteristic of a great earthquake. Of significant interest in this slide is the great distance from the epicenter to which this earthquake was felt. It is said that the quake rang church bells in New England, more than 1000 miles away.



Isoseismal Map for 1886 Charleston Earthquake

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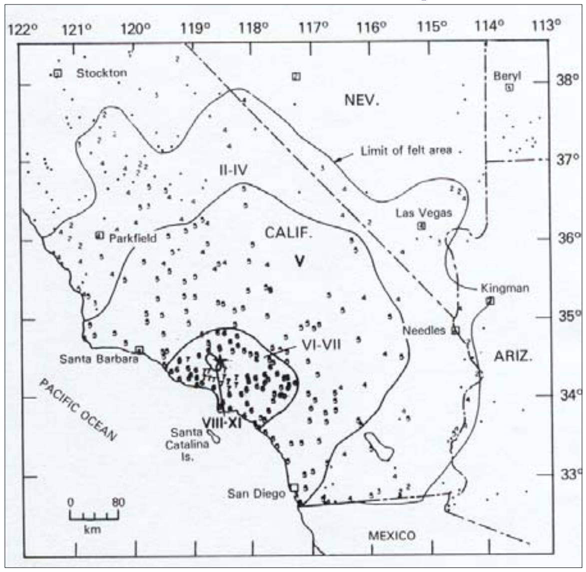
For the 1886 Charleston earthquake, the epicentral intensity was X. Note that the earthquake was felt as far away as Chicago (MMI = V) and New York City (MMI = IV). As with the New Madrid earthquake, the felt distance was quite extensive.

Isoseismal Map for February 9, 1971, San Fernando Earthquake



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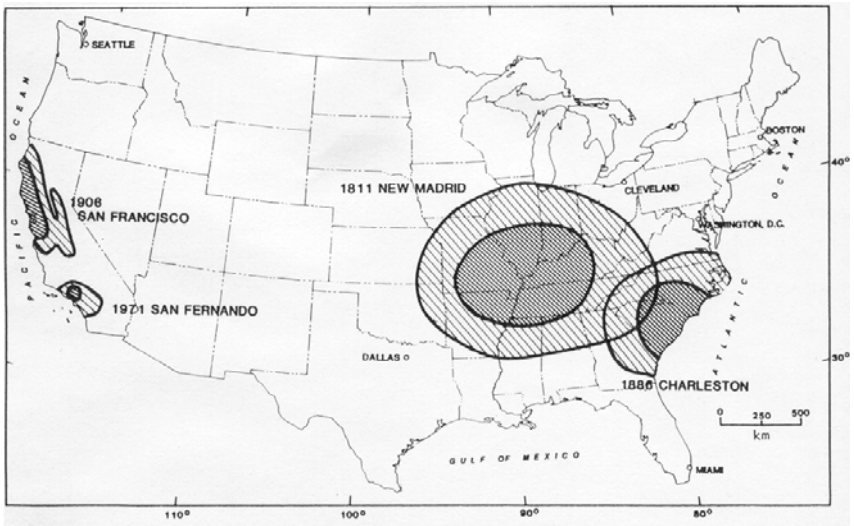
This isoseismal map from the 1971 earthquake in the San Fernando Valley of California shows intensities (here shown as Arabic numerals at the individual assessment points). The maximum intensity is in the range of XI, similar to the New Madrid earthquake and close to that of the Charleston earthquake (I = X). Of significant interest, however, is the range of the felt area resulting from this California earthquake -- on the order of a 100 to 200 miles. This is in stark contrast to the eastern quakes that had felt areas of more than 1000 miles.

Comparison of Isosiesmal Intensity for Four Earthquakes



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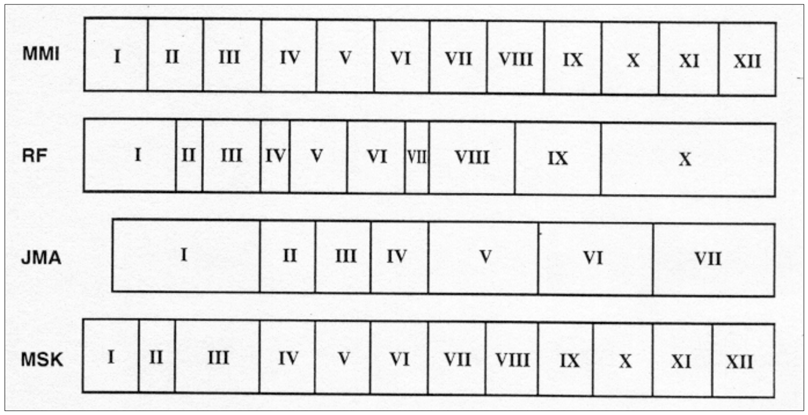
This map shows isoseismal maps for several (nonconcurrent) earthquakes. The minimum MMI value shown on the maps is approximately VI (boundaries of felt regions would be significantly greater). As noted earlier, the extent of the isoseismal boundary VI is much greater in the eastern United States than in the western states. This is because the crustal region of the West is located near a plate boundary and is therefore much more internally fractured and less homogenous than the relatively less fractured East. A good analogy is a bell. An uncracked bell will ring much more clearly and loudly than a cracked one. The fall off of intensity with distance is called *attenuation*. Because of the local differences, several different attenuation relationships have been developed for the United States. Attenuation relationships are an important element of seismic hazard analysis and are described in some detail under that topic.

Comparisons of Various Intensity Scales



MMI = Modified Mercalli
RF = Rossi-Forel
JMA = Japan Meteorological Agency
MSK =Medvedez-Spoonheur-Karnik

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There are other intensity measures that may produce a quite different number (e.g., the Japan Meteorological Society scale which goes up to only VII) for the same earthquake so users of intensity should be careful when using an intensity scale and particularly when attempting to convert intensity to magnitude for the purpose of seismic hazard analysis.

Instrumental Seismicity

Magnitude (Richter, 1935)

Also called local magnitude

M_L = Log [Maxumum Wave Amplitude (in mm/1000)]

Recorded Wood-Anderson seismograph

100 km from epicenter

The most recognized instrument measure of an earthquake is the *Richter* or *local magnitude*. A wave with a 1-mm amplitude would produce a Richter Magnitude 3, and a 1-meter amplitude would be Richter 6. Because it is unlikely that an instrument will be placed exactly 100 km from the epicenter, correction factors may be used to adjust for other epicentral distances. The Richter scale is limited to shallow crustal California-type earthquakes and saturates (defined later) at about Richter Magnitude 7. The Richter scale does not distinguish between different types of waves. Because of the limited usefulness of the Richter scale, other magnitude measures have been developed.

Magnitude (in general)

$$M = Log\ A + f(d,h) + C_S + C_R$$

A is wave amplitude

$F(d,h)$ accounts for focal distance and depth

C_S and C_R, are station and regional corrections

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This slide provides a bit more detail regarding magnitude measurements.

Other Wave-Based Magnitudes

M_S Surface-wave magnitude (Rayleigh waves)

m_b Body-wave magnitude (P waves)

M_B Body-wave magnitude (P and other waves)

m_{bLg} (Higher order Love and Rayleigh waves)

M_{JMA} (Japanese, long period)

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These are some of the other magnitude measures in use. They are basically similar to the Richter but measure different types of waves.

Moment Magnitude

Seismic moment = $M_O = \mu AD$

[Units = force times distance]

Where:
μ = modulus of rigidity
A = fault rupture area
D = fault dislocation or slip

Moment magnitude = $M_W = (\mathrm{Log}\ M_O - 16.05)/1.5$

(Units = dyne-cm)

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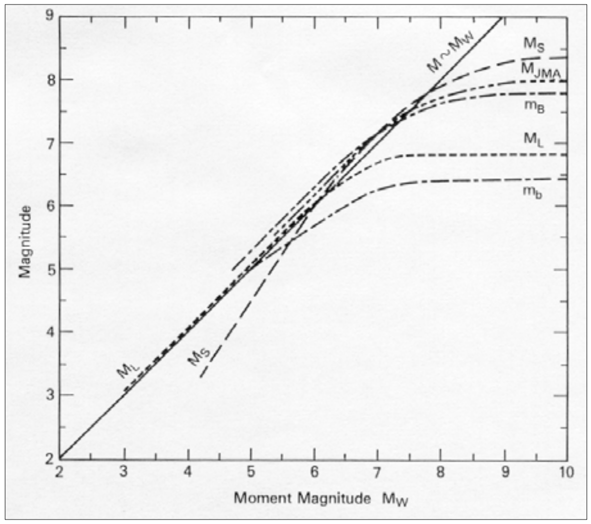
Because of the different types of magnitude scale used, it is necessary to have a "common denominator" scale. Such a scale, called *moment magnitude*, is based on a quantity called the *seismic moment*. The moment may be determined for actual earthquakes and may be estimated for quakes yet to occur. Note that the seismic moment has units of energy and that the moment magnitude is the log of the energy.

Note, if Mo=1E25, Mw=6.

If Mo=1E28, Mw=8.

Hence, for a 1000 time increase in energy, magnitude goes up by 2. For 31.6 time energy, magnitude goes up by 1. [Suggestion: have class do this exercise]

Moment Magnitude vs Other Magnitude Scales (Magnitude Saturation)



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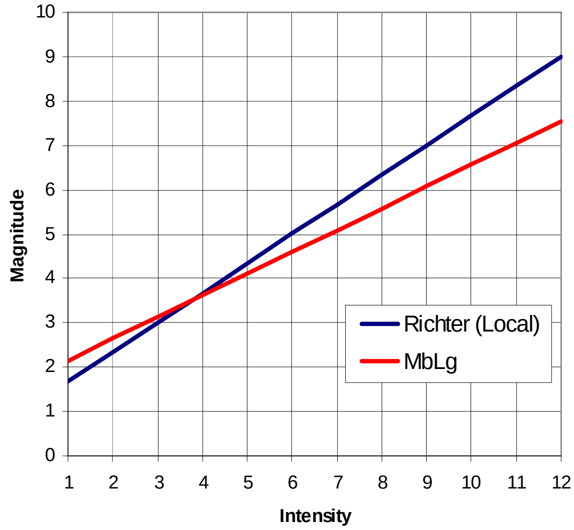
This slide shows how the different magnitude scales relate to moment magnitude. As may be seen, some of the magnitude scales *saturate* at about Magnitude 6.5 to 7. Hence, a Richter-based seismometer will only record a Magnitude 6.8 earthquake when the true (moment) magnitude may be as large as 9. For example, the 1960 Chile earthquake had M_s = 8.3, but M_w = 9.5. This does not mean Richter is no good but rather that it should only be used to measure shallow crustal earthquakes. It is not likely that the Earth's crust has enough strength to generate earthquakes with magnitude greater than about M_w = 9.5.

Approximate Relationship Between Magnitude and Intensity



$$\hat{M}_L = 0.67 I_0 + 1$$

$$\hat{m}_{bLg} = 0.49 I_0 + 1.66$$

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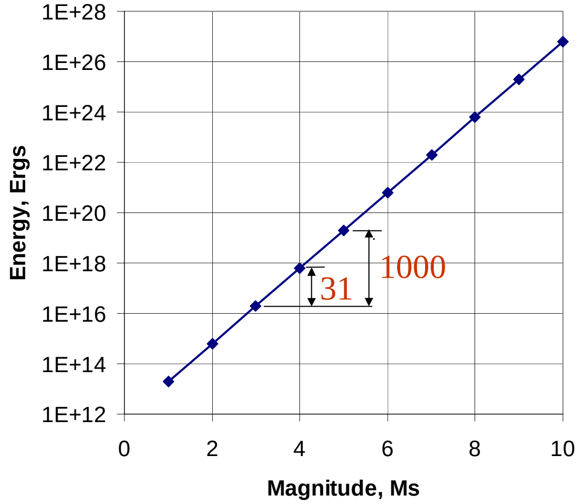
For the purpose of seismic hazard analysis (where magnitude is required), it is necessary to derive empirical relationships between intensity and magnitude and then to use this relationship to estimate magnitude when only intensity (e.g., New Madrid) is available. Note that a different intensity-to-magnitude relationship is required for each magnitude type.

Note also that these empirical relationships relate magnitude to epicentral intensity. This can be a dubious transformation because the highest intensity does not necessarily occur at the epicenter.

Seismic Energy Release

$$\text{Log } E = 1.5\, M_S + 11.8$$



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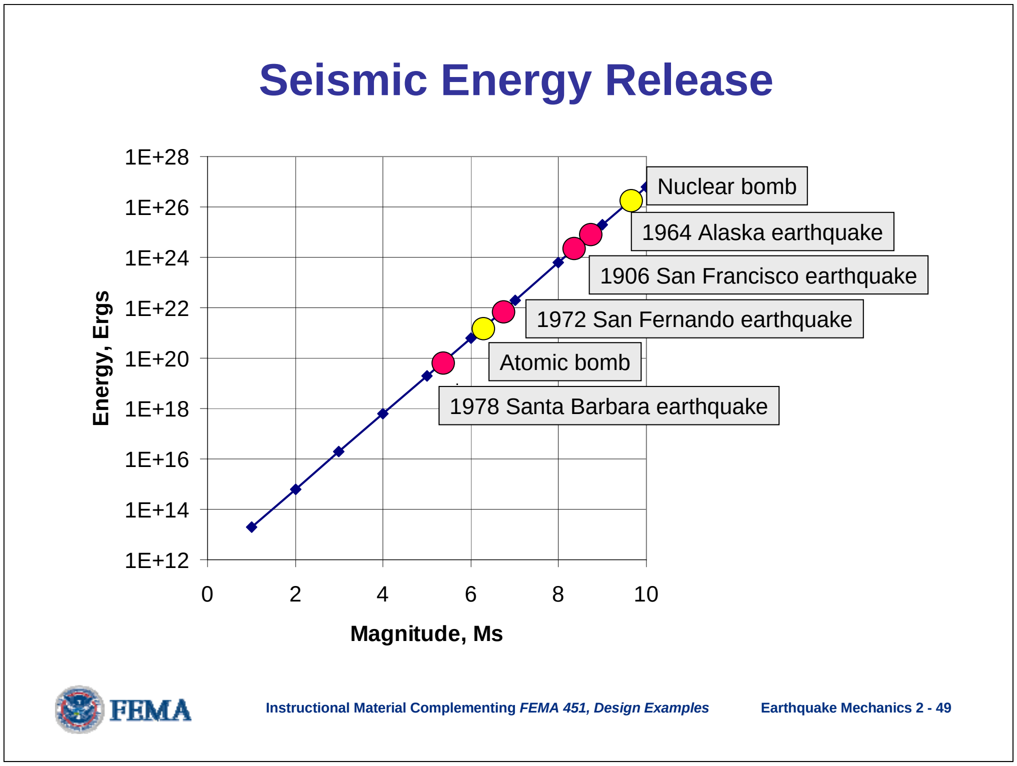
This is the relationship between energy and magnitude. It repeats the information presented earlier.



This slide shows, in a relative sense, how the energy released by a earthquake compares to atomic and nuclear bombs. Clearly, the damage caused by the 1972 San Fernando Valley earthquake cannot be compared to that caused by an atomic bomb (e.g., Hiroshima). It is important to recognize, however, that even by 1972 building standards, the quality of seismic resistance in Southern California was reasonably good. Compare this to the earthquakes in Armenia, Turkey, and India where the devastation was certainly equivalent to that caused by the dropping of an atomic bomb -- almost complete destruction and loss of tens of thousands of lives.

Seismic building standards have improved significantly since 1972, and it should therefore be expected that if the 1972 earthquake was repeated today, relatively less damage and loss of life would occur. However, there are always lessons to be learned, witness the unexpected damage to steel moment frames during the Northridge earthquake.

Ground Motion Accelerograms

Sources:

- NONLIN (more than 100 records)
- Internet (e.g., National Strong Motion Data Center)
- USGS CD ROM

Uses:

- Evaluation of earthquake characteristics
- Development of response spectra
- Time history analysis

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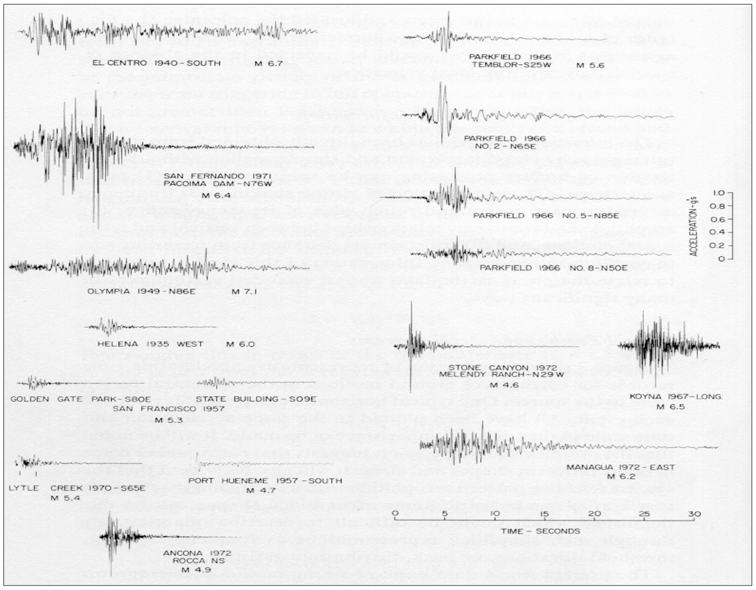
This slide introduces a second way to directly measure earthquakes at a site distant from the epicenter. Here, ground *accelerations* are measured directly. These accelerations may give a measure of the size of the actual earthquake, but they are not as useful as indicators of intensity and magnitude. Nevertheless, experienced seismologists can learn a great deal about the nature of the earthquake by studying the accelerogram and other entities derived from the accelerogram.

Note that websites provided by such entities as the U.S. Geological Survey provide earthquake information just a few hours after earthquakes occur; prior to the availability of the internet, it could take months to obtain records, particularly from foreign earthquakes.

Sample Ground Motion Records



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This slide shows a variety of recorded horizontal ground accelerations. Recordings are generally made in three mutually orthogonal directions: X, Y (horizontal) and vertical. The X and Y orientations are often, but not always, aligned with N-S or E-W compass directions. In a sense, the accelerogram is the “fingerprint” of the earthquake. No two are exactly alike even when generated from different earthquakes on the same fault. Note the vast differences in duration and frequency for the earthquakes shown. Note also that all quakes are to the same scale.

Ground Motion Characteristics

- Acceleration, velocity, displacement
- Effective peak acceleration and velocity
- Fourier amplitude spectra
- Duration (bracketed duration)
- Incremental velocity (killer pulse)
- Response spectra
- Other (see, for example, Naiem and Anderson 2002)

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These are some of the characteristics of recorded ground motion.

Corrected vs Uncorrected Motions

Corrections made primarily:

- To remove instrument response
- To account for base line shift

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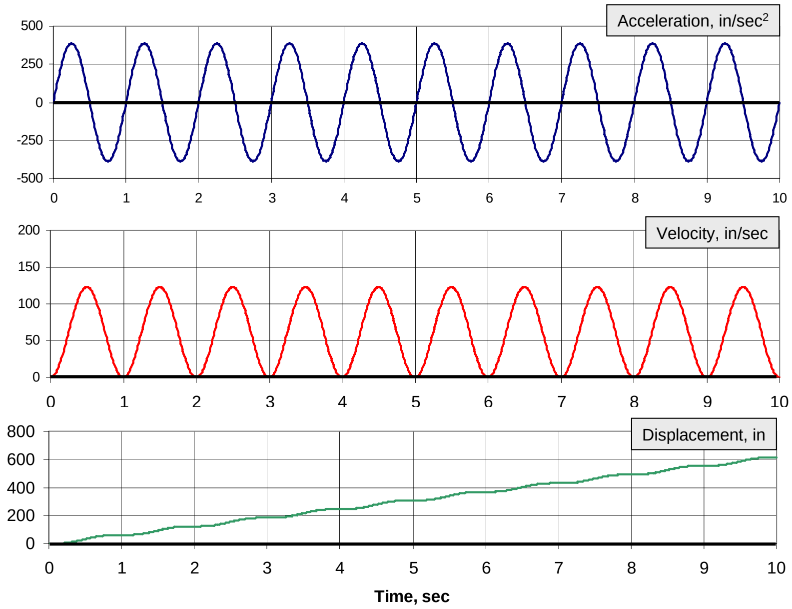
Uncorrected accelerograms are not directly useful in structural analysis unless they are "corrected" to remove instrument response and to account for base line shift. When retrieving motions from the standard databases, it is usually very clear whether the motions have been corrected in the description of the motion.

Base Line Correction for Simple Ground Motion



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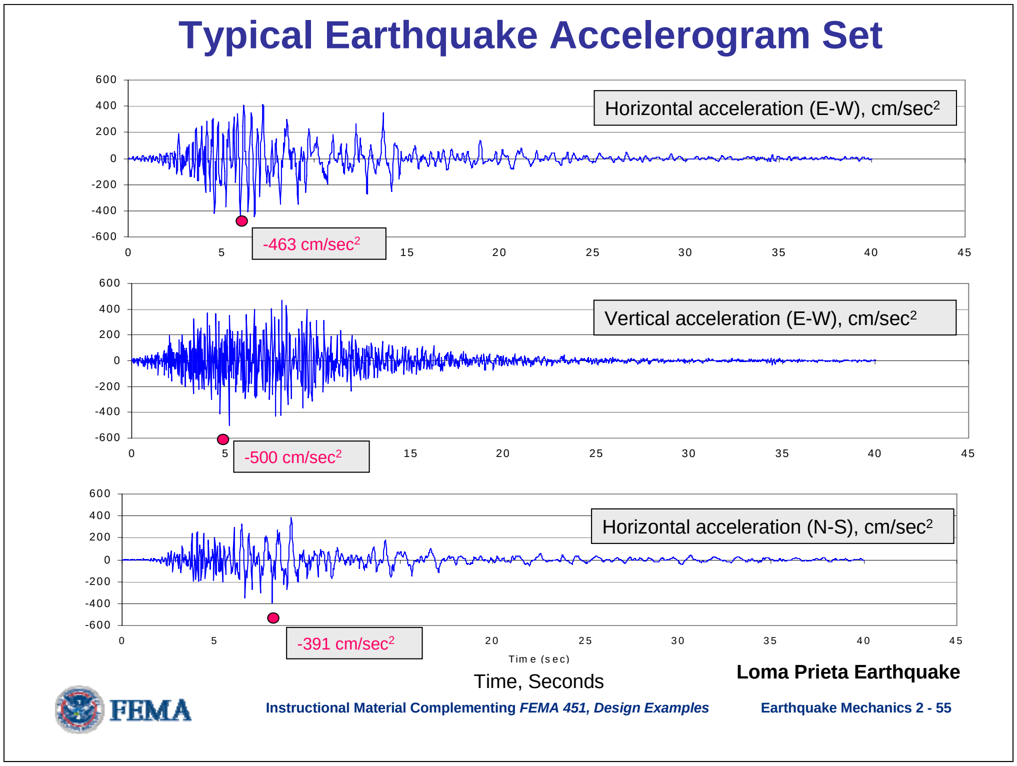
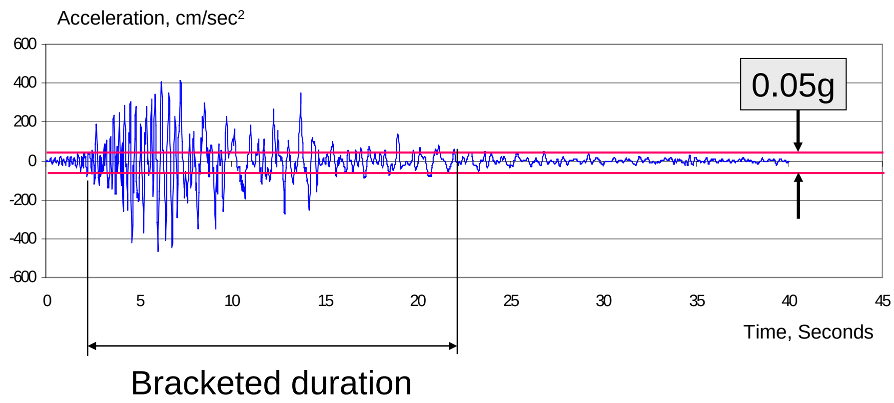
It is often desirable to numerically integrate ground motions to obtain velocity and displacement records. Unfortunately, the constants of integration are not known and the result may be a severe base line drift, even for the sine wave shown.

Most ground motion analysts use a linear regression analysis to correct the base line -- first for velocity and then again for displacement. What is known for sure is that at the end of the earthquake, the velocity should be zero. However, the displacement at the end of the earthquake may not be zero as some residual ground displacement may occur.



This is a recording of the corrected ground motions occurring during the Loma Prieta earthquake, which occurred on October 19, 1989, with an epicenter about 100 km south of Oakland, California. The earthquake had an epicentral intensity of VIII and a surface magnitude M_s of 7.1. (Intensities in San Francisco and parts of Oakland were higher (IX) due to site amplification effects.) The earthquake had a peak horizontal and vertical acceleration of about 0.5g. Note the difference in apparent frequency content between the middle (vertical) and the top and bottom (horizontal) records. Strong shaking lasted only about 20 seconds but that was long enough to take down several important structures.

Definition of Bracketed Duration



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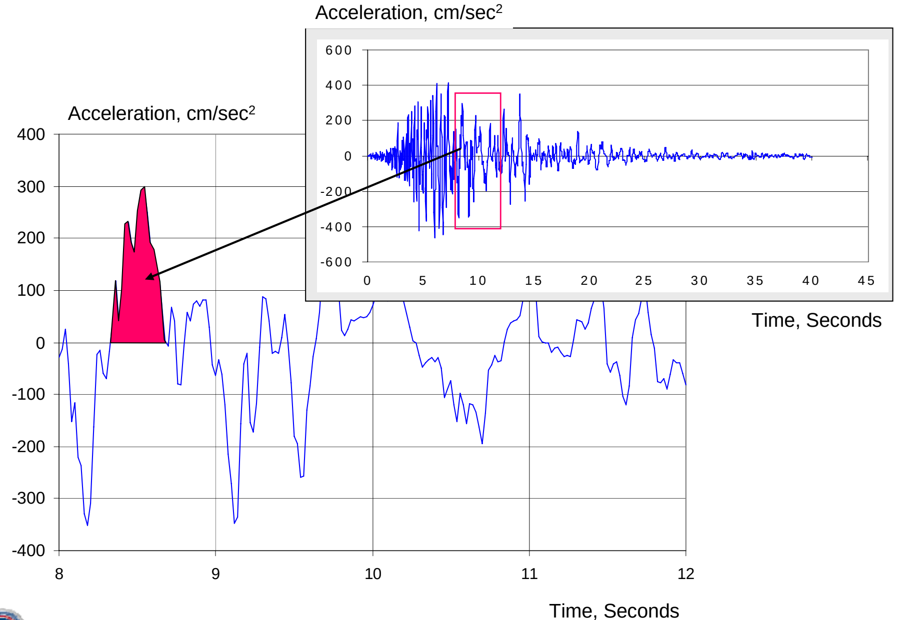
Some researchers such as Naiem and Anderson refer to the "bracketed duration" of a ground motion. This is defined as the duration between the first and last occurrence of an acceleration exceeding 0.05g. For structures responding inelastically to earthquakes, the bracketed duration is an important measure of damage potential.

Definition of Incremental Velocity



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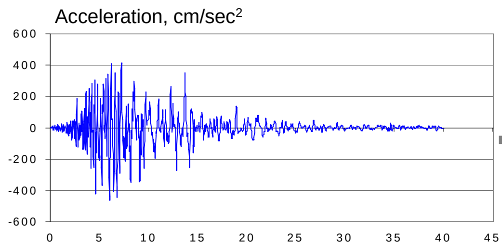
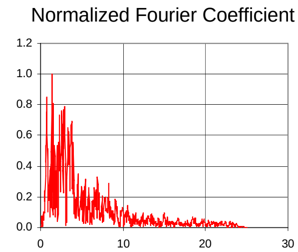
Another measure of damage is incremental velocity, which is defined as the area of individual acceleration pulses. It has been demonstrated that ground motions with high incremental velocities may be extremely damaging. High incremental velocities are characteristic of "near source" ground motions, particularly when the seismic rupture front is propagating towards the site.

Concept of Fourier Amplitude Spectra

$$\ddot{v}_g(t) \cong a_0 + \sum_{j=1}^{N/2} a_j \cos(2\pi j f_0) + \sum_{j=1}^{N/2} b_j \sin(2\pi j f_0) = a_0 + \sum_{j=1}^{N/2} A_j \cos(2\pi j f_0 + \phi_j)$$

$$f_0 = df = 1/Ndt \qquad \phi_j = \arctan\left(-\frac{b_j}{a_j}\right) \qquad A_j = \sqrt{a_j^2 + b_j^2}$$





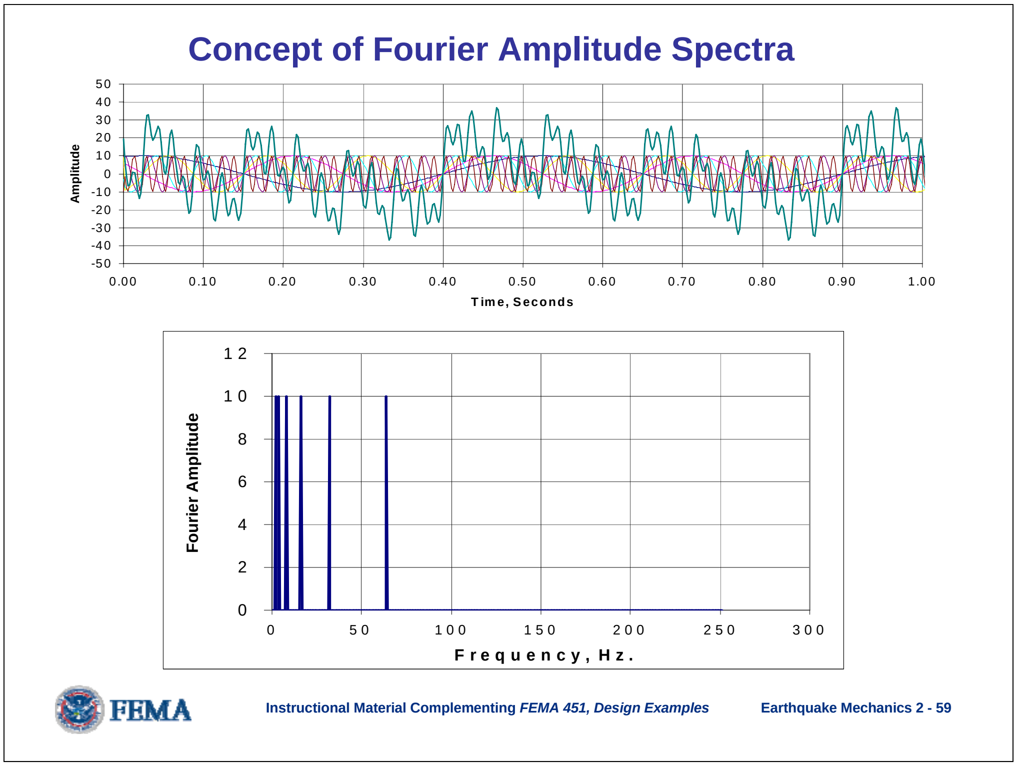
N points at timestep *dt*

N/2 points at frequency *df*

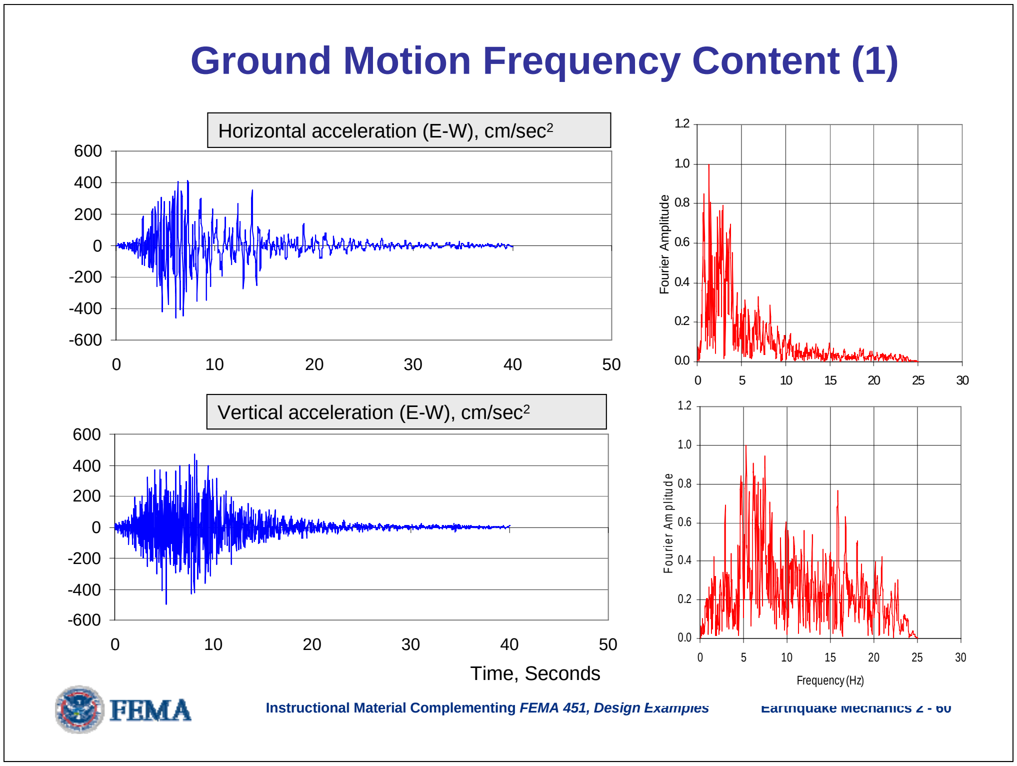
Given a corrected ground motion, there are two important tools useful for analyzing the motion: the Fourier amplitude spectra and the linear elastic response spectra.

The Fourier amplitude spectra is a plot of the result of a fast Fourier transform (FFT) of the original record.

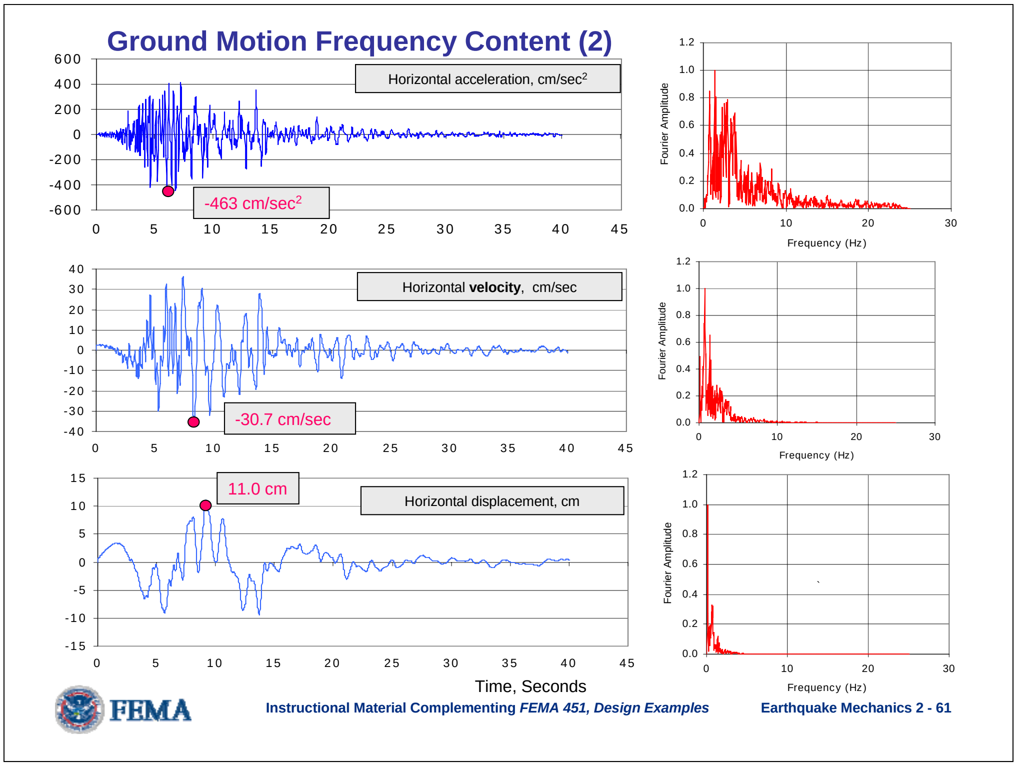
Example: For a record with 1024 points recorded at a time step of 0.01 seconds, the maximum recoverable frequency (Nyquist frequency) is 1/2 of 1/0.01 or 50 Hz. The FFT will give a real and imaginary amplitude at 512 different frequencies at a frequency interval of 50/512 or 0.0975 Hz. The amplitude spectrum is a plot of the SRSS of the imaginary and real amplitude as computed at each frequency interval.



This is an example of a Fourier amplitude spectra of an artificial motion that consists of the superposition of five sine waves of constant amplitude. The green line is the composite of the five individual components. In this case, the Fourier spectra consists of five spikes, each occurring at the appropriate period and amplitude.

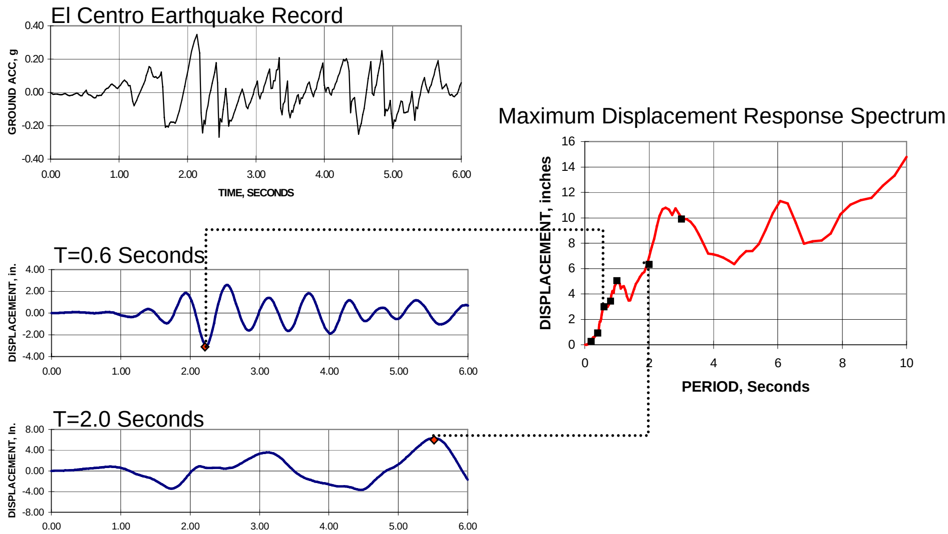


This slide shows the Fourier amplitude spectrum of a the Loma Prieta earthquake. Different spectra are given for the horizontal (top) and vertical (bottom) components of motion. Note that in comparison to the horizontal motion, the vertical motion is much richer in higher frequencies and relatively weaker in the lower frequencies.



Here, Fourier amplitude spectra are given for the horizontal acceleration, velocity, and displacement records of the Loma Prieta earthquake. Note that the higher frequencies attenuate much more rapidly for velocity and particularly for displacement. The reason for this is clear from the mathematics as the acceleration amplitude is omega squared times the displacement amplitude for a simple sine wave. (Omega = circular frequency = 2 pi times cyclic frequency.) The high frequency content is present in the velocity and displacement but it just has a much lower amplitude relative to the lower frequencies.

Development of an Elastic Displacement Response Spectrum



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Another measure of the effect of the acceleration record is the elastic response spectrum. This type of spectrum is described in detail in the topic on single-degree-of-freedom dynamics. The response spectrum is also integral with the seismic hazard maps as explained in the topic on seismic hazard analysis.