LECTURE 22: EARTHQUAKE SEISMOLOGY

The seismic cycle consists of the slow buildup of elastic strain within rocks until the stresses within the rock exceed the strength of the rock. At this point, the rock fails, and the accumulated strain energy is released within a few seconds during an earthquake.

Earthquakes typically occur on a fault that slips an average distance $D$ over a rupture area $S$. The first patch to slip is called the focus, or hypocenter, and is located at a focal depth below a surface point known as the epicenter.

The distance between an epicenter and a seismic station is the epicentral distance, and can be determined from the difference in time between the arrival of the P-waves and S-waves. The epicenter and origin time of an earthquake can be determined if the epicentral distance is known for 3 seismic stations.

For local earthquakes travel time is a linear function of distance $D$: 
\[ t_p - t_0 = D/\alpha \text{ for P-waves} \quad \text{and} \quad t_s - t_0 = D/\beta \text{ for S-waves} \]

where \( t_p \) and \( t_s \) are the P- and S- arrival times, and \( t_0 \) is the origin time. Then:

\[ t_s - t_p = D \left( \frac{1}{\beta} \frac{1}{\alpha} \right) = (t_p - t_0) \left( \frac{\alpha}{\beta} - 1 \right) \]

For large epicentral distances, the curved ray path through the earth shortens the travel time, but \( t_s - t_s \) still uniquely determines \( D \). If the earthquake is located at depth, then the epicentral distance \( \Delta \) may be smaller than \( D \). In this case the focal depth can be approximated as: \( d = \sqrt{D^2 - \Delta^2} \).

By studying the first motions from seismographs surrounding an epicenter, an earthquake's sense of slip (its focal mechanism) can be determined:

If the first motion of an arriving P-wave is upward, then slip on the fault must have been toward the location of that station. The sense of first motion from many stations plotted on a stereonet yields a pattern of compression and extension from which two perpendicular fault planes (active and auxiliary) can be determined. Other (tectonic or aftershock) data must be used to determine which of these faults planes is the active one. The type of fault (thrust, normal, strike-slip, or an intermediate type) is defined by these “beachball” diagrams.
The amplitude of P-waves transmitted from a fault displacement varies as:
\[ A(r, t, \alpha, \theta) = A_0(r, t, \alpha) \sin^2(2\theta) \]
and thus has a quadrupole character (4 lobes). The axes of the maximum tensional (T) and compressional (P) deviatoric stresses that are present before the earthquake are aligned with those of the maximum first-motion compressive or extensive ground motions produced by the earthquake.

Notably, nuclear explosions exhibit compression in all directions.

The **intensity** of an earthquake is a subjective measure of the earthquake’s visible effects. Variations depend on distance from the earthquake, focusing of earthquake energy, and building designs, and are used to assess seismic risk.

The **magnitude** of an earthquake is determined instrumentally. When measured from surface waves:
\[ M_s = \log_{10}\left( \frac{A_S}{T} \right) + 1.66 \log_{10}(\Delta) + 3.3 \]
where \( A_S \) is the vertical ground motion (in \( \mu m \)) of the maximum Rayleigh-wave amplitude, \( T \) is the period (18-22 s) and \( 20^\circ \leq \Delta \leq 160^\circ \) is the epicentral distance. When measured from body waves:
\[ m_b = \log_{10}\left( \frac{A_P}{T} \right) + Q(\Delta, h) \]
where \( Q \) is an empirical correction for attenuation.
Body wave amplitude \( A_p \) is the max amplitude of P-waves with period \( T < 3 \) s. Magnitude saturation is reached for \( M_s \sim 8 \) and \( M_b \sim 6 \). The **moment magnitude** avoids magnitude saturation for large earthquakes: 
\[
M_w = \frac{2}{3} \left( \log_{10}(M_0) - 9.1 \right)
\]
where \( M_0 \) is the **seismic moment** in N m.

The seismic moment is a measurement of the slip along the fault as given by:
\[
M_0 = \mu S D \text{ where } S \text{ & } D \text{ are rupture area & slip, and } \mu \text{ is the rigidity modulus}
\]

The seismic moment can be determined by inverting for the source parameters of an earthquake using synthetic with observed seismograms.

The seismic moment can be empirically related to maximum intensity using:
\[
I_{\text{max}} = 1.5M_s - 1.8 \log_{10} h + 1.7 \text{ where } h \text{ is focal depth}
\]

Seismic moment can be empirically related to an earthquake’s energy release:
\[
\log_{10} E = 4.8 + 1.5M_s - 1.2 + 2.4m_b \text{ where } E \text{ is energy release in joules.}
\]

Thus \( 1000 = 10^{(1.5*2)} \) earthquakes of \( M_s=6 \) release the energy of one \( M_s=8 \) event.

The **Gutenberg-Richter relation** expresses the annual number earthquakes with magnitudes exceeding \( M_S \):
\[
\log_{10} N = a - bM_S
\]
where \( b \sim 1 \) for global seismicity.

The recurrence time for an earthquake of a given magnitude is given by \( 1/N \).

The largest earthquake expected annually is \( a \) if \( b=1 \) (because \( \log_{10} 1 = 0 \)).