Workshop on Megathrust Earthquakes and Tsunami

Prof. Thorne Lay

CONCEPTS:

Basic nature of Waves
P waves and S waves
Seismic Rays and Wavefronts
Seismometers
Travel Time Curves

Supplementary Reading (Optional, for more details/rigor)
Lay and Wallace, *Modern Global Seismology*, Ch. 2
Stein and Wysession, *An Introduction to Seismology, Earthquakes and Earth Structure*, Ch. 2

One solution is for both sides to equal zero all the time:

The scalar potential side:

$$\nabla^2 \phi(\mathbf{x}, t) = \frac{1}{\alpha^2} \frac{\partial^2 \phi(\mathbf{x}, t)}{\partial t^2}$$

with velocity $\alpha = [(\lambda + 2\mu)/\rho]^{1/2}$

α is the P velocity, and it depends on the two Lamé parameters λ, μ, or on incompressibility K, μ, along with density, ρ.

The vector potential side:

$$\boldsymbol{\nabla}^2 \boldsymbol{\Upsilon}(\mathbf{x}, t) = \frac{1}{\beta^2} \frac{\partial^2 \boldsymbol{\Upsilon}(\mathbf{x}, t)}{\partial t^2}$$

with velocity $\beta = (\mu/\rho)^{1/2}$

β is the S velocity, and it depends on only μ, and ρ.

For a Poisson medium which has equal Lamé parameters, $\alpha = 1.732\ \beta$.

The general mathematical representation of motions everywhere in the medium satisfying the elastodynamic equations, $\mathbf{U}(\mathbf{x},t)$, is given by simple spatial derivative operations on two space-time functions, $\phi(\mathbf{x},t)$ and $\psi(\mathbf{x},t)$, which are themselves solutions of the three-dimensional wave equation.

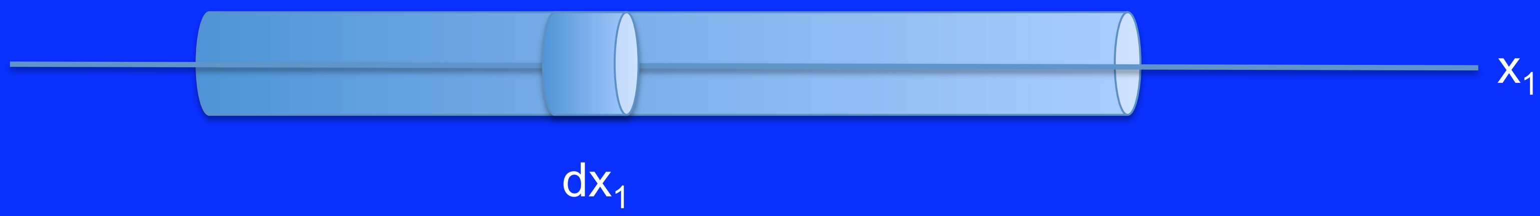
$\phi(\mathbf{x},t)$ is the P wavefield
$\psi(\mathbf{x},t)$ is the S wavefield

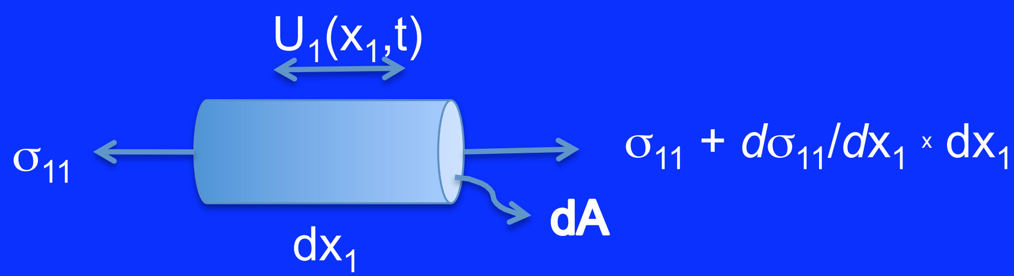
Physical displacements of the P wave are calculated by taking the gradient of $\phi(\mathbf{x},t)$.

Physical displacements of the S wave are calculated by taking the curl of $\psi(\mathbf{x},t)$.

Because the wavefields satisfy the wave equation, P and S wave motions have basic behavior of waves – this makes the final solution very straightforward: we just need to understand properties of waves.

Thin, elastic rod; all motion in the x_1 direction.





Mass = density × volume = $\rho\, dA\, dx_1$

Equation of motion : sum forces = Mass × acceleration

$$\rho\, dx_1 dA \times d^2u_1/dt^2 = (\sigma_{11} + d\sigma_{11}/dx_1 \times dx_1)\, dA - \sigma_{11}\, dA$$

$\rho\, d^2u_1/dt^2 = d\sigma_{11}/dx_1$ Spatial gradient of stress balanced by inertial terms

Need constitutive law, measure in lab: $\sigma_{11} = E\, \varepsilon_{11}$, E= Young's modulus

$\sigma_{11} = E\ \varepsilon_{11} = E\ du_1/dx_1$

$\rho\ d^2u_1/dt^2 = d\sigma_{11}/dx_1 = E\ d^2u_1/d^2x_1$: now have equation of motion all in terms of u_1

Let $c = (E/\rho)^{1/2}$

$d^2u_1/dt^2 = c^2\ d^2u_1/d^2x_1$ We finally arrive at the one-dimensional wave equation.

The implication is that the displacements in the rod will spread along it as a wave, obeying the wave equation.

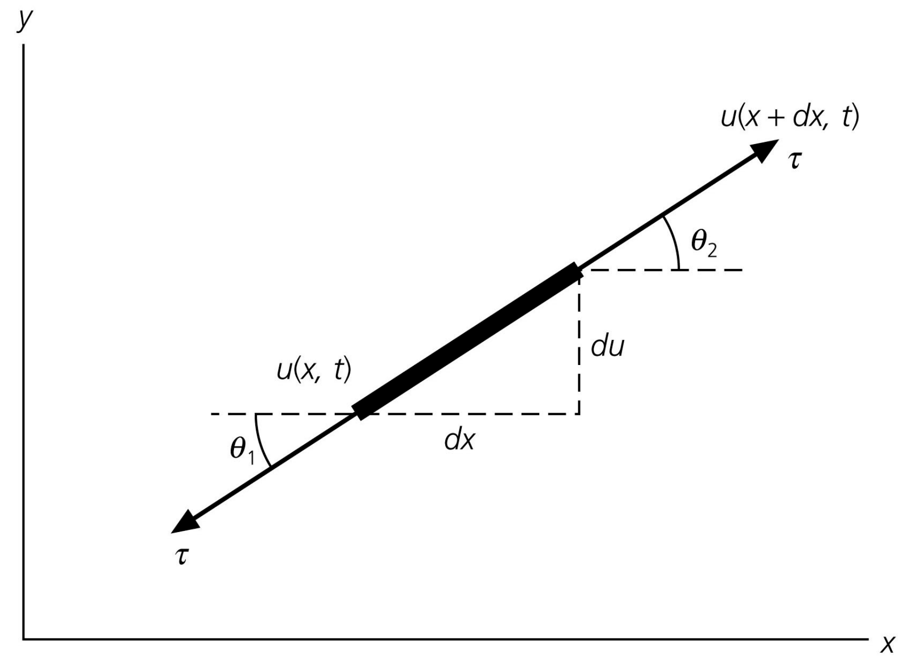
Note – this one-dimensional treatment is not physically correct; for a finite colume rod, there is more than 1 strain term produced by σ_{11}, and the full 3D solution before is needed for rigorous solution, and it would give eise to two waves in the rod (P and S waves), as a general solution to elastic waves in a solid. But, we see the parallel development easily for this simple example. The same holds for the string problem, next.

$\mathbf{F} = m\mathbf{a}$

OK, let's quickly derive a simple 1D wave equation from a standard physics problem: motion of a string.

Tension along/within the string is constant, τ

Figure 2.2-1: Tensions on a string segment.



$$\mathbf{F} = m\mathbf{a}$$

$$F(x, t) = \tau \sin\theta_2 - \tau \sin\theta_1 = \rho dx \frac{\partial^2 u(x, t)}{\partial t^2}$$

The sum of the forces in the y-direction equals the mass times acceleration in the y-direction. We use linear density, ρ, such That ρdx is mass.

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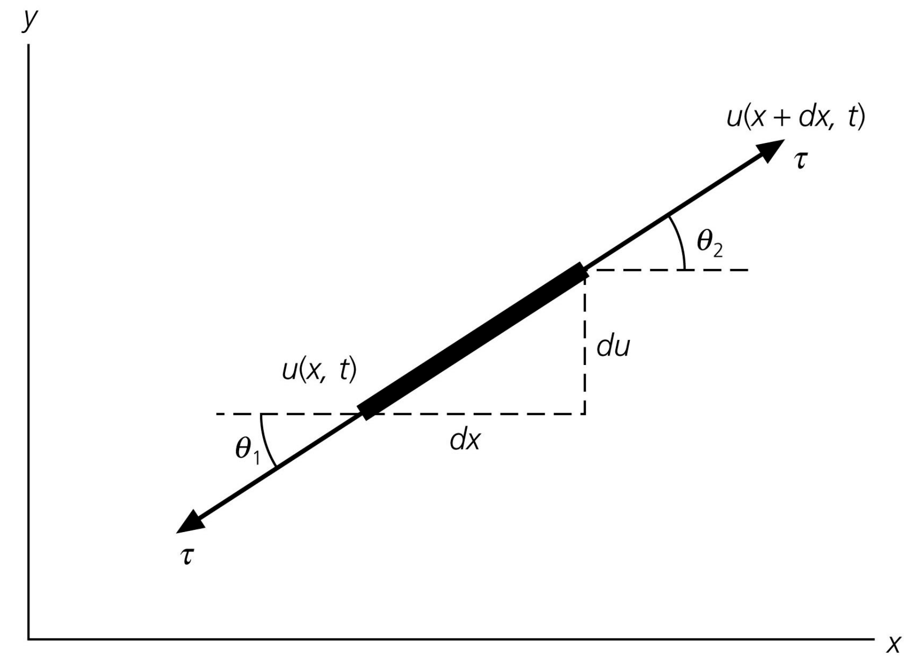

$$\mathbf{F} = m\mathbf{a}$$

$$F(x,\, t) = \tau \sin\theta_2 - \tau \sin\theta_1 = \rho dx \,\frac{\partial^2 u(x,\, t)}{\partial t^2}$$

For θ small, $\sin\theta \approx \theta \approx \tan\theta$:

$$\tau\left(\frac{\partial u(x + dx,\, t)}{\partial x} - \frac{\partial u(x,\, t)}{\partial x}\right) = \rho dx \,\frac{\partial^2 u(x,\, t)}{\partial t^2}$$

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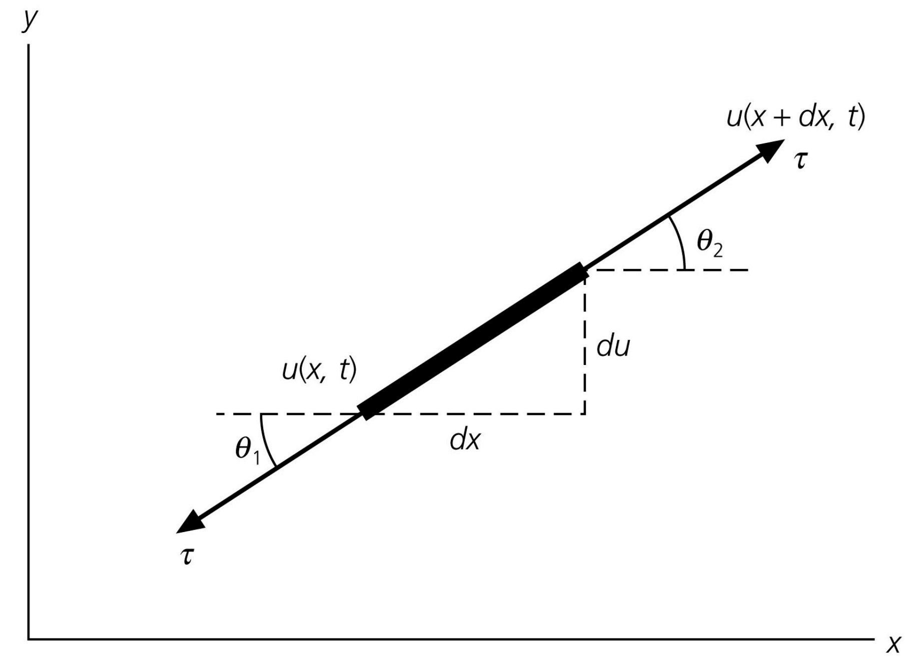

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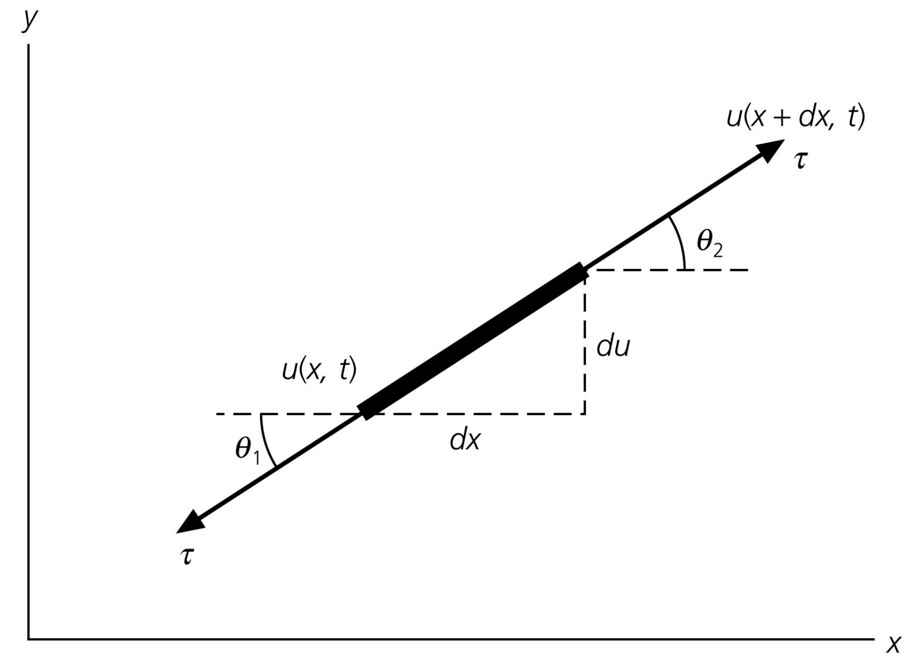

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Discarding the higher Taylor series terms:

$$\tau\left(\frac{\partial u(x,\, t)}{\partial x} + \frac{\partial^2 u(x,\, t)}{\partial x^2}\, dx - \frac{\partial u(x,\, t)}{\partial x}\right) = \tau\, \frac{\partial^2 u(x,\, t)}{\partial x^2}\, dx = \rho dx\, \frac{\partial^2 u(x,\, t)}{\partial t^2}$$

$\mathbf{F} = m\mathbf{a}$

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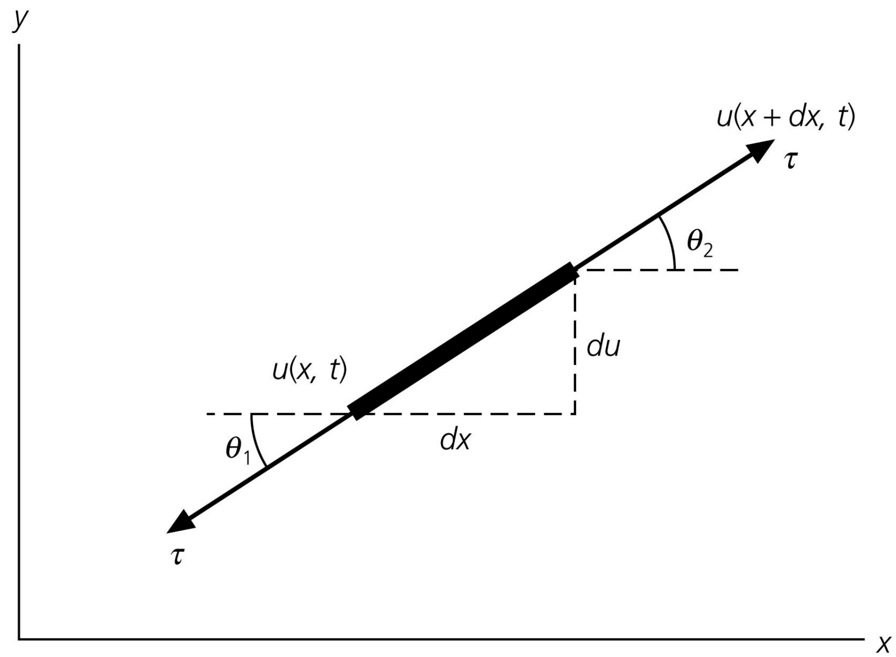

$$F(x, t) = \tau \sin\theta_2 - \tau \sin\theta_1 = \rho dx \frac{\partial^2 u(x, t)}{\partial t^2}$$

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Wave equation:

$$\frac{\partial^2 u(x, t)}{\partial x^2} = \frac{1}{\mathrm{v}^2} \frac{\partial^2 u(x, t)}{\partial t^2}$$

where $\mathrm{v} = (\tau/\rho)^{1/2}$

NOTE: The spatial imbalance in forces gives accelerations. The wave equation always has two spatial derivatives and two temporal derivatives – this controls the nature of solutions.

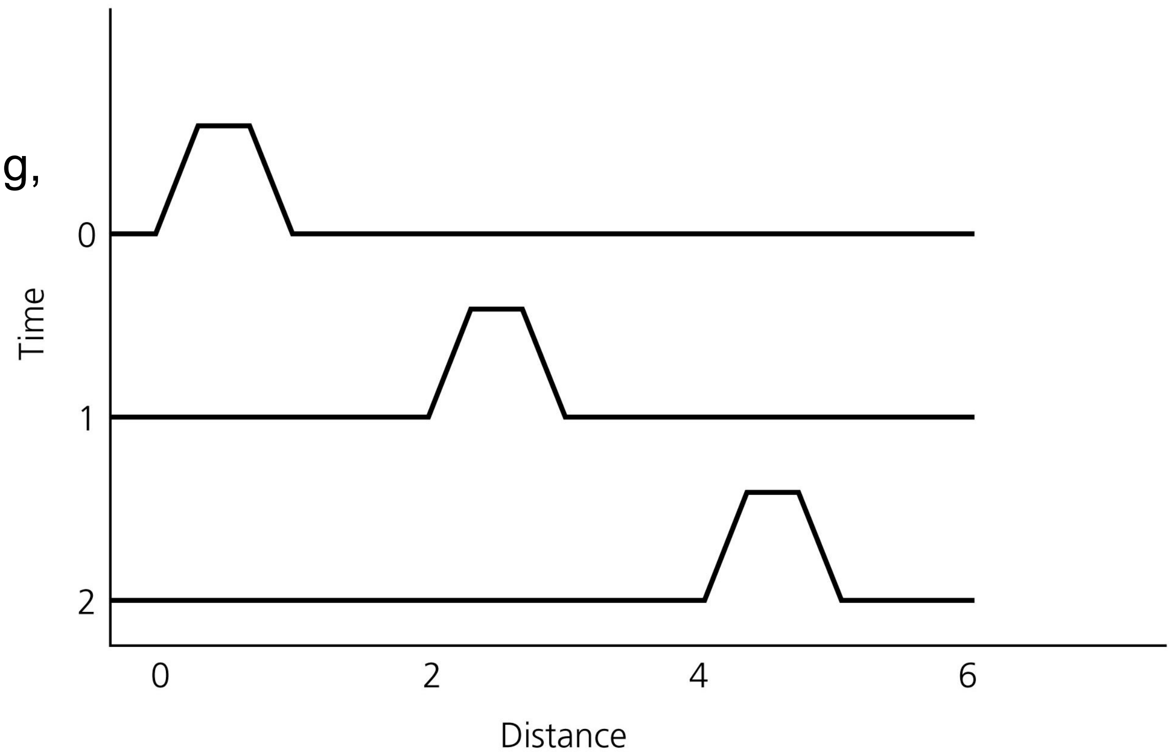
The wave equation is a hyperbolic partial differential equation, with two space and two time derivatives of the solution weighted by a relative weighting factor which is the inverse square of the wave velocity.

This is a classic PDE, and applied mathematicians have figured out all properties of the solutions for different choice of coordinate system (Cartesian, spherical, cylindrical).

The most important point is that the solutions (waves) are disturbances with a given shape (functional form) that retains that exact shape as space and/or time vary. The connection between the space and time domains that keep the waveshape fixed is given by the wave velocity (distance/time).

Essentially, whatever shape the P wave or S wave has generated by the source will translate through the medium as a function of time without distortion. This enables analysis of seismic sources using wave motions recorded thousands of miles away!

Figure 2.2-2: Propagating pulse, *f(x - 2t)*.



For our string; if we put an initial arbitrary shape on the string, that shape will move through space as a function of time, with a velocity controlled by the physical parameters of the medium.

$$\frac{\partial^2 u(x,t)}{\partial x^2} = \frac{1}{v^2}\frac{\partial^2 u(x,t)}{\partial t^2}$$

The solution has the form $u(x, t) = f(x \pm vt)$:

Here, f is ANY function, of an argument involving space (x) and time (t) with a weighting factor of wave velocity (v), in the form of x ± vt. The chain rule for two derivatives with respect to x and two derivatives with respect to t ensure the wave equation will be satisfied.

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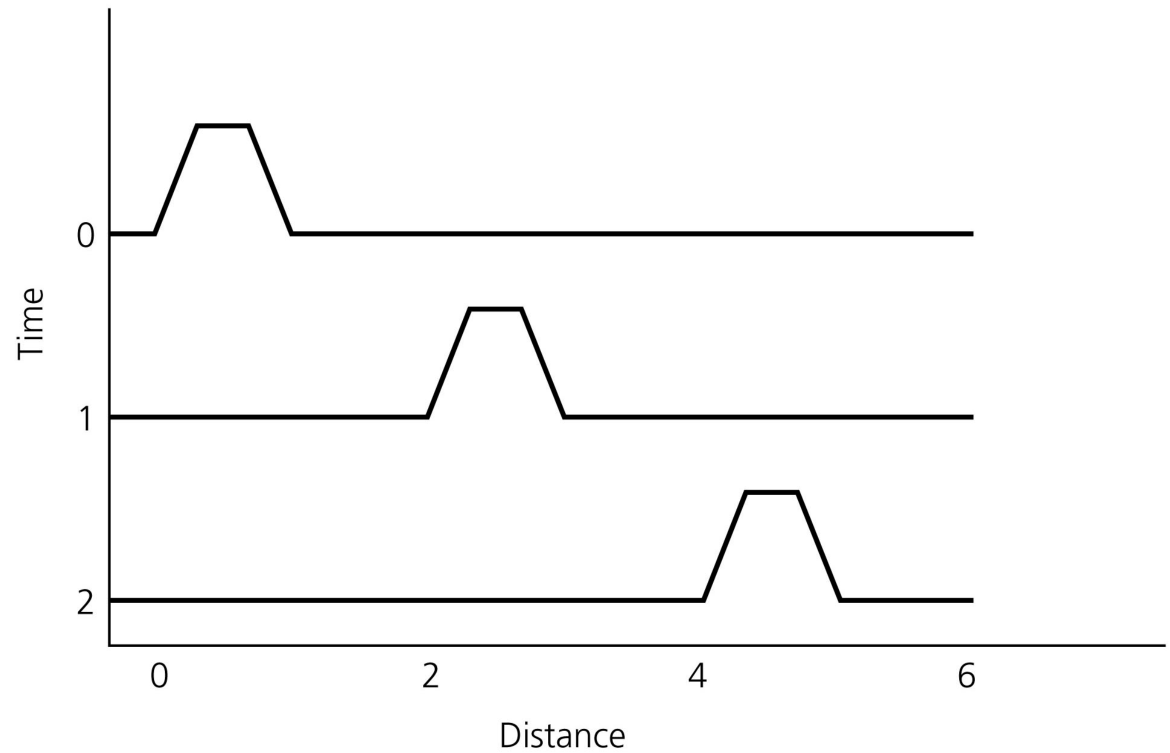

$$\frac{\partial^2 u(x,\, t)}{\partial x^2} = \frac{1}{\mathrm{v}^2}\,\frac{\partial^2 u(x,\, t)}{\partial t^2}$$

The solution has the form $u(x,\, t) = f(x \pm \mathrm{v}t)$:

$$\frac{\partial^2 u(x,\, t)}{\partial x^2} = f''(x \pm \mathrm{v}t) \qquad \text{and} \qquad \frac{\partial^2 u(x,\, t)}{\partial t^2} = \mathrm{v}^2 f''(x \pm \mathrm{v}t)$$

where f'' is the second derivative of f

One solution is for both sides to equal zero all the time:

The scalar potential side:

$$\nabla^2 \phi(\mathbf{x}, t) = \frac{1}{\alpha^2} \frac{\partial^2 \phi(\mathbf{x}, t)}{\partial t^2}$$

with velocity $\alpha = [(\lambda + 2\mu)/\rho]^{1/2}$

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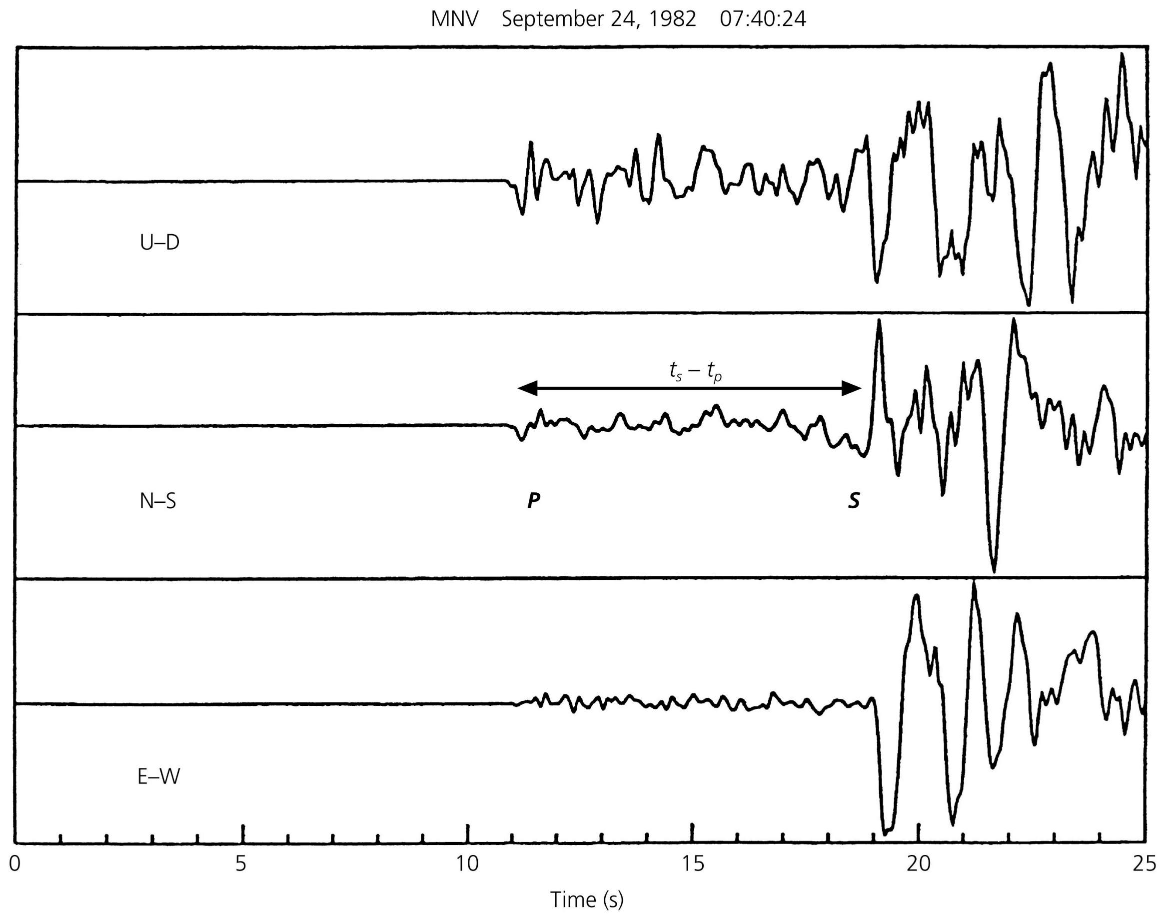
For a Poisson medium which has equal Lamé parameters, α = 1.732 β.

Figure 2.4-9: Seismograms recorded 64 km from a small, shallow earthquake.



For close distances (horizontal propagation):

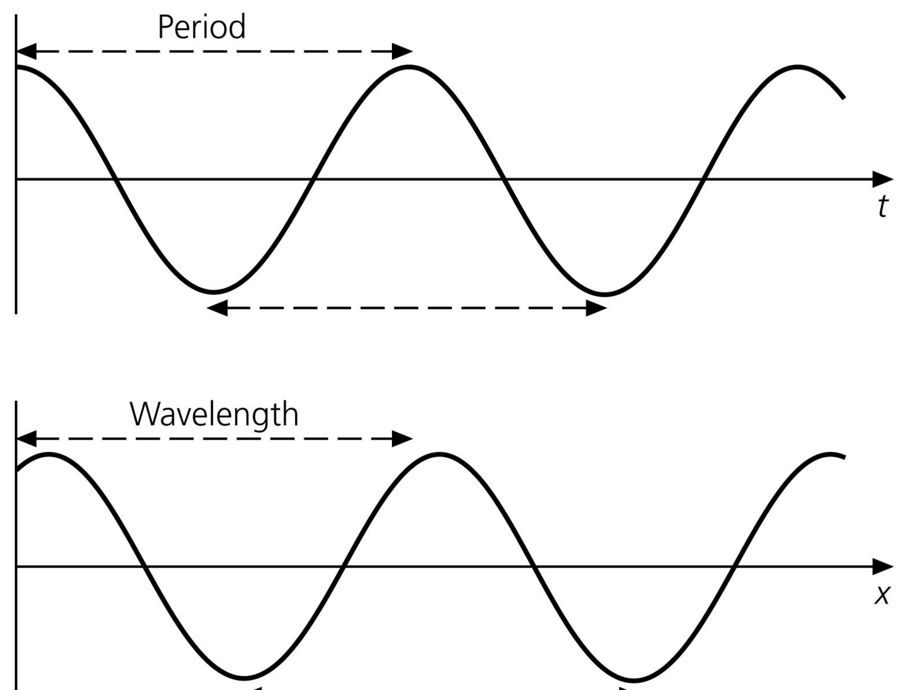
$$t_s = x/3.2 \qquad t_p = x/5.5$$

$$t_s - t_p = x(1.0/3.2 - 1.0/5.5) = x/7.6$$

Table 2.2-1 Relationships Between Wave Variables

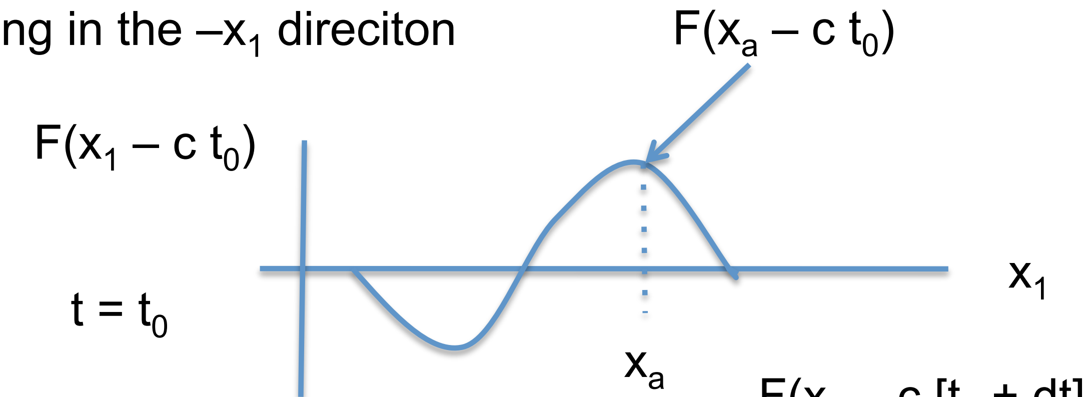
QUANTITY	UNITS	
Velocity	distance/time	$\mathrm{v} = \omega/k = f\lambda = \lambda/T$
Period	time	$T = 2\pi/\omega = 1/f = \lambda/\mathrm{v}$
Angular Frequency	time^{-1}	$\omega = 2\pi/T = 2\pi f = k\mathrm{v}$
Frequency	time^{-1}	$f = \omega/(2\pi) = 1/T = \mathrm{v}/\lambda$
Wavelength	distance	$\lambda = 2\pi/k = \mathrm{v}/f = \mathrm{v}T$
Wavenumber	distance^{-1}	$k = 2\pi/\lambda = \omega/\mathrm{v} = 2\pi f/\mathrm{v}$

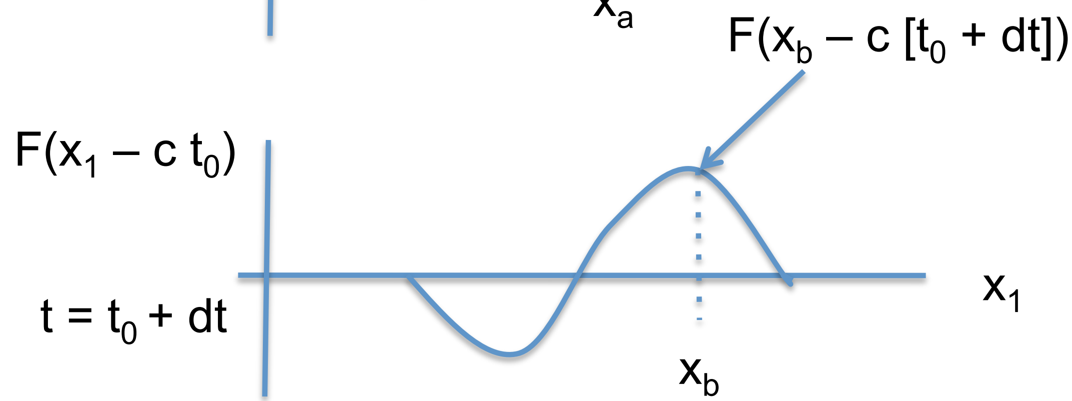
Figure 2.2-4: Harmonic wave, $u = A \cos(\omega t - kx)$.



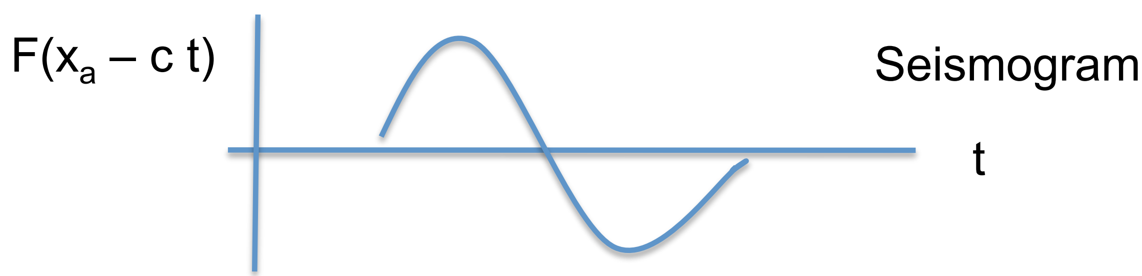
$F(x_1 - c\ t)$: wave traveling in the $+x_1$ direction
$G(x_1 + c\ t)$: wave traveling in the $-x_1$ direciton

Consider $F(x_1 - c\ t)$





$x_b - x_a = c\ dt$
$c = dx/dt =$ velocity



Consider $F(x_a - ct)$

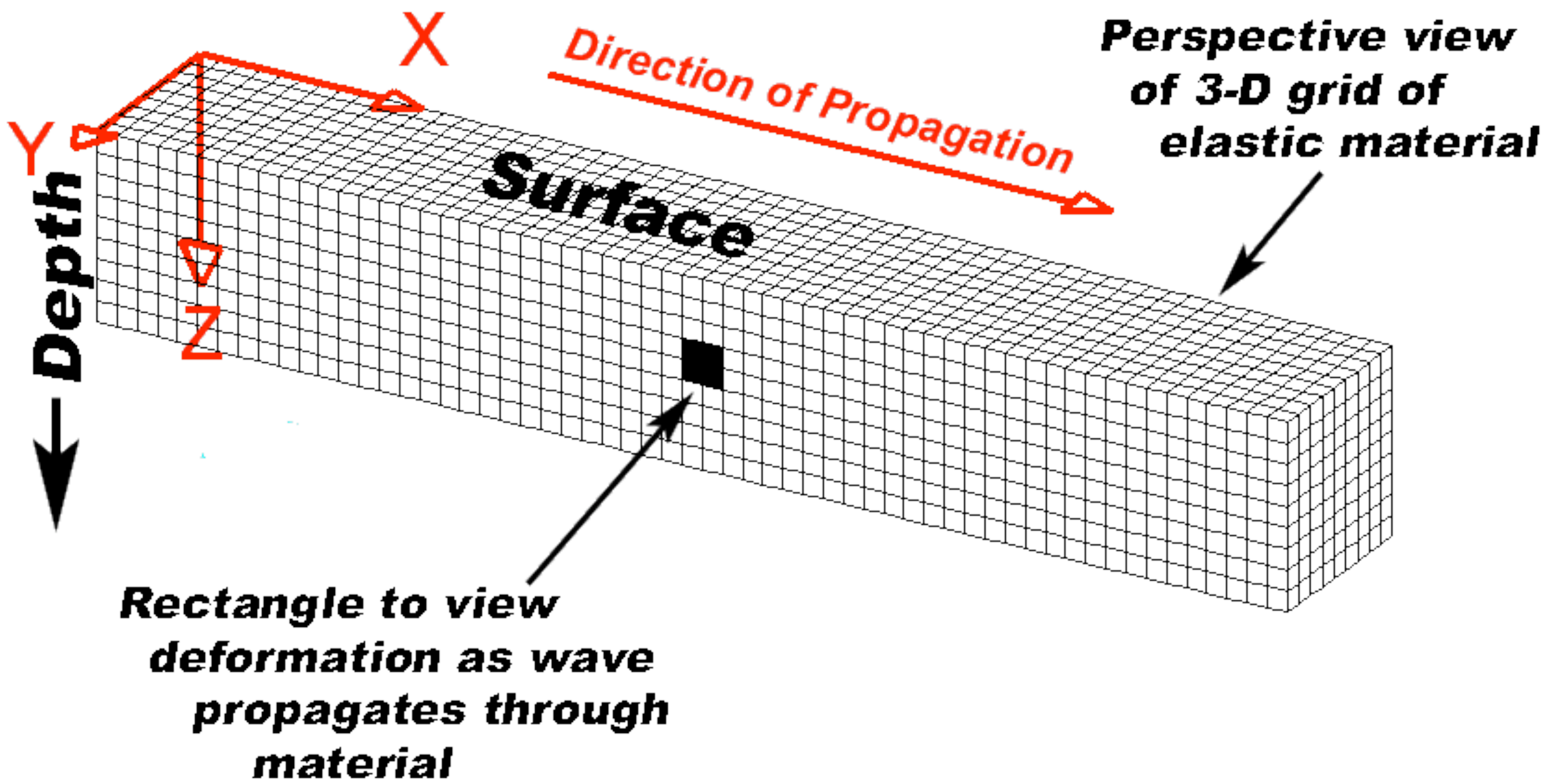
The wave equation is a hyperbolic partial differential equation, with two space and two time derivatives of the solution weighted by a relative weighting factor which is the inverse square of the wave velocity.

This is a classic PDE, and applied mathematicians have figured out ALL properties of the solutions for different choices of coordinate system (e.g., Cartesian, spherical, cylindrical).

The most important point is that the solutions (waves) are disturbances with a given shape (functional form) that retains that exact shape as space and/or time vary with a specific scale factor. The connection between the space and time domains that keep the waveshape fixed is given by the wave velocity (distance/time).

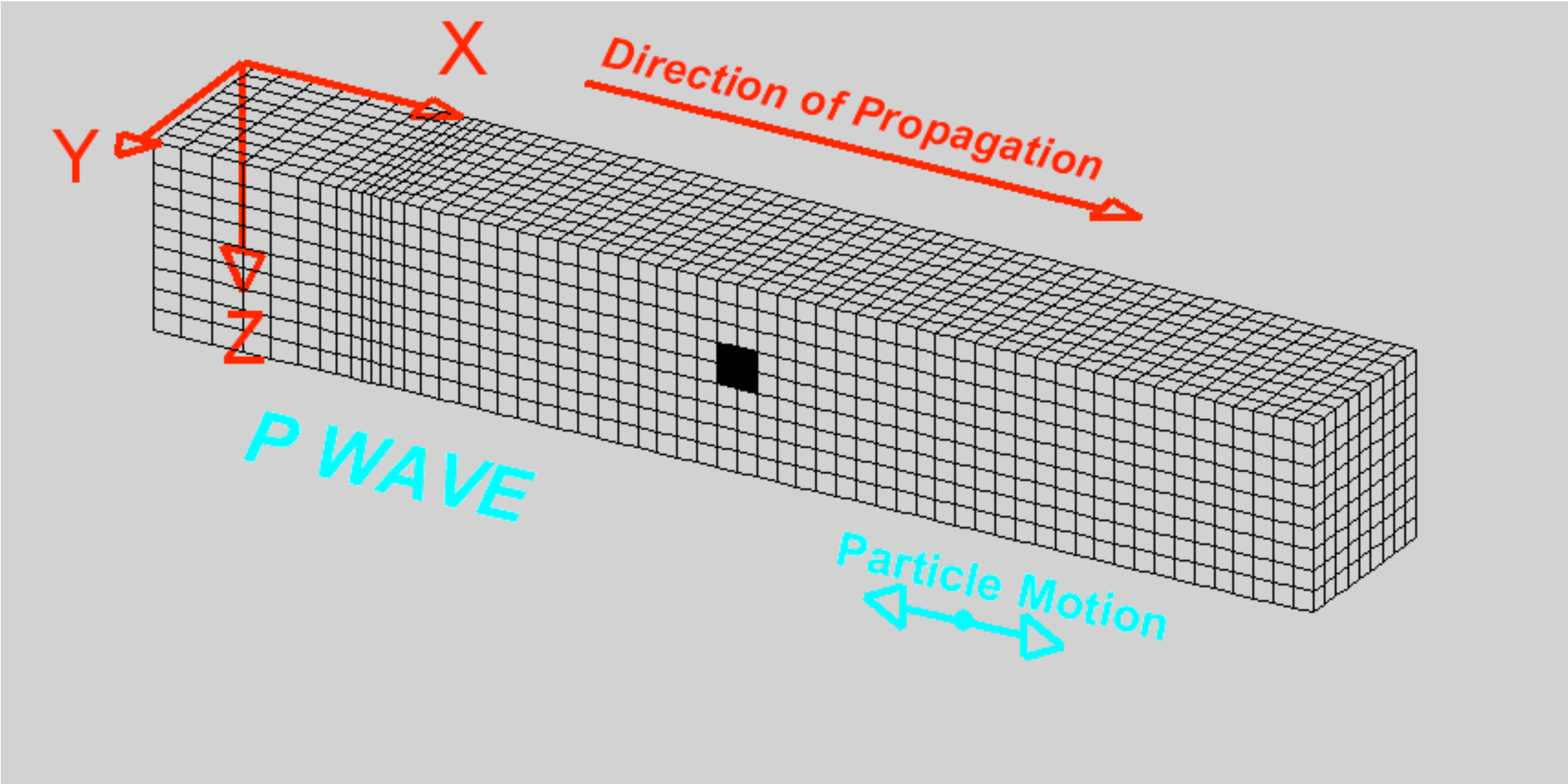
Essentially, whatever shape the P wave or S wave has generated by the source will translate through the medium as a function of time without distortion at a velocity determined by the physical properties of the medium. This enables analysis of seismic sources using wave motions recorded thousands of miles away!

3-D Grid for Seismic Wave Animations



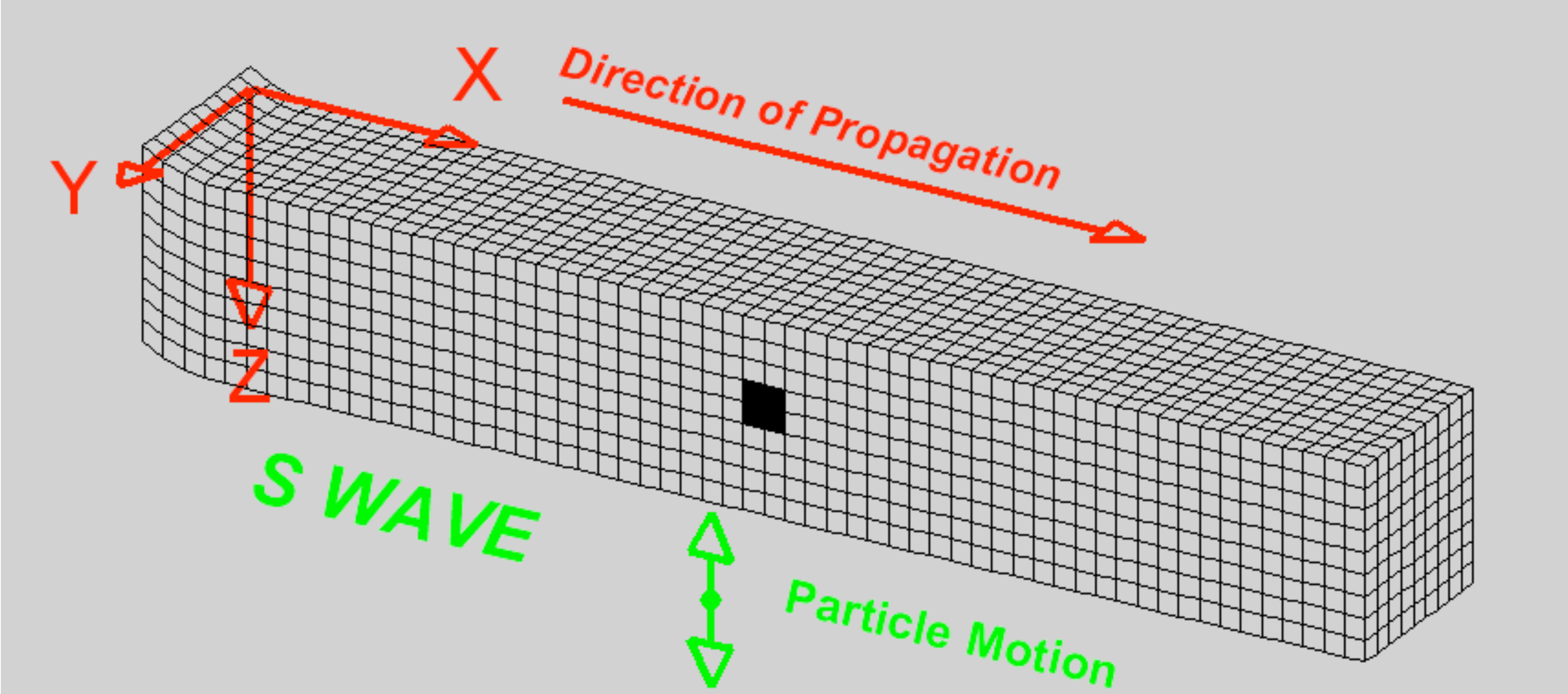
No *attenuation* (decrease in amplitude with distance due to spreading out of the waves or absorption of energy by the material) *dispersion* (variation in velocity with frequency), or *anisotropy* (velocity depends on direction of propagation) is included.

Compressional Wave (P-Wave) Animation



Deformation propagates. Particle motion consists of alternating compression and dilation. Particle motion is parallel to the direction of propagation (longitudinal). Material returns to its original shape after wave passes.

Shear Wave (S-Wave) Animation



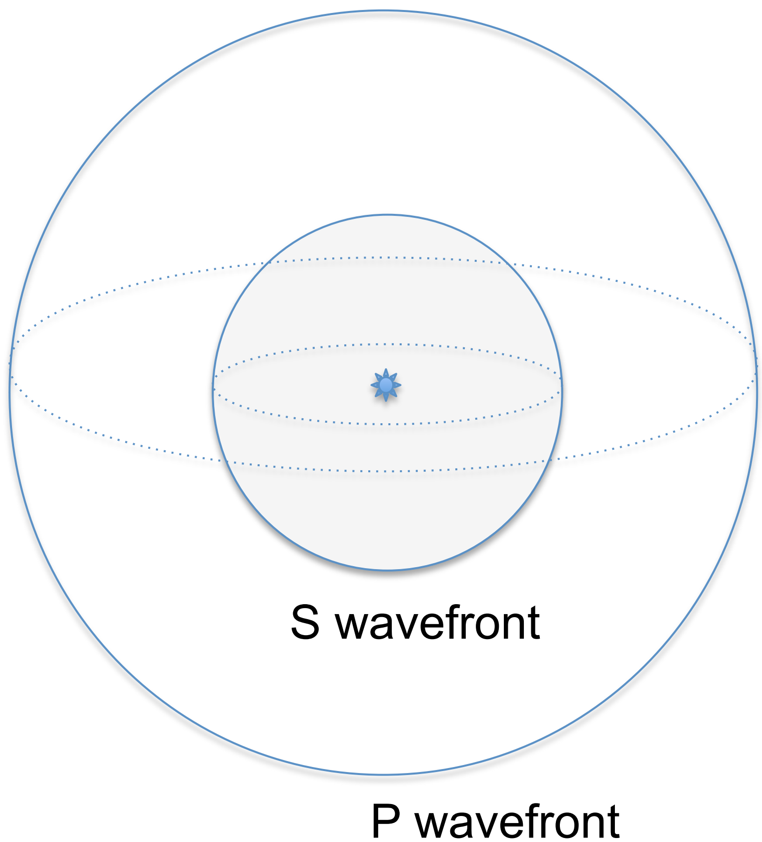
Deformation propagates. Particle motion consists of alternating transverse motion. Particle motion is perpendicular to the direction of propagation (transverse). Transverse particle motion shown here is vertical but can be in any direction. However, Earth's layers tend to cause mostly vertical (SV; in the vertical plane) or horizontal (SH) shear motions. Material returns to its original shape after wave passes.

Seismic Body Waves

Wave Type (and names)	Particle Motion	Other Characteristics
P, **Compressional**, Primary, Longitudinal	Alternating compressions ("pushes") and dilations ("pulls") which are directed in the same direction as the wave is propagating (along the raypath); and therefore, perpendicular to the wavefront.	P motion travels fastest in materials, so the P-wave is the first-arriving energy on a seismogram. Generally smaller and higher frequency than the S and Surface-waves. P waves in a liquid or gas are pressure waves, including sound waves.
S, **Shear**, Secondary, Transverse	Alternating transverse motions (perpendicular to the direction of propagation, and the raypath); commonly approximately polarized such that particle motion is in vertical or horizontal planes.	S-waves do not travel through fluids, so do not exist in Earth's outer core (inferred to be primarily liquid iron) or in air or water or molten rock (magma). S waves travel slower than P waves in a solid and, therefore, arrive after the P wave.

Seismic waves spread in all three directions from a source energy release.

At a given instant in time t_1, the motions generated by a source at time t_0, will have expanded out as spherical waves. There will be a P wave and an S wave, with the P wave having traveled further from the source than the S wave because it has higher velocity.



P velocity $V_p = [(K + 4/3\mu)/\rho]^{1/2}$

S velocity $V_s = [\mu/\rho]^{1/2}$

K – incompressibility

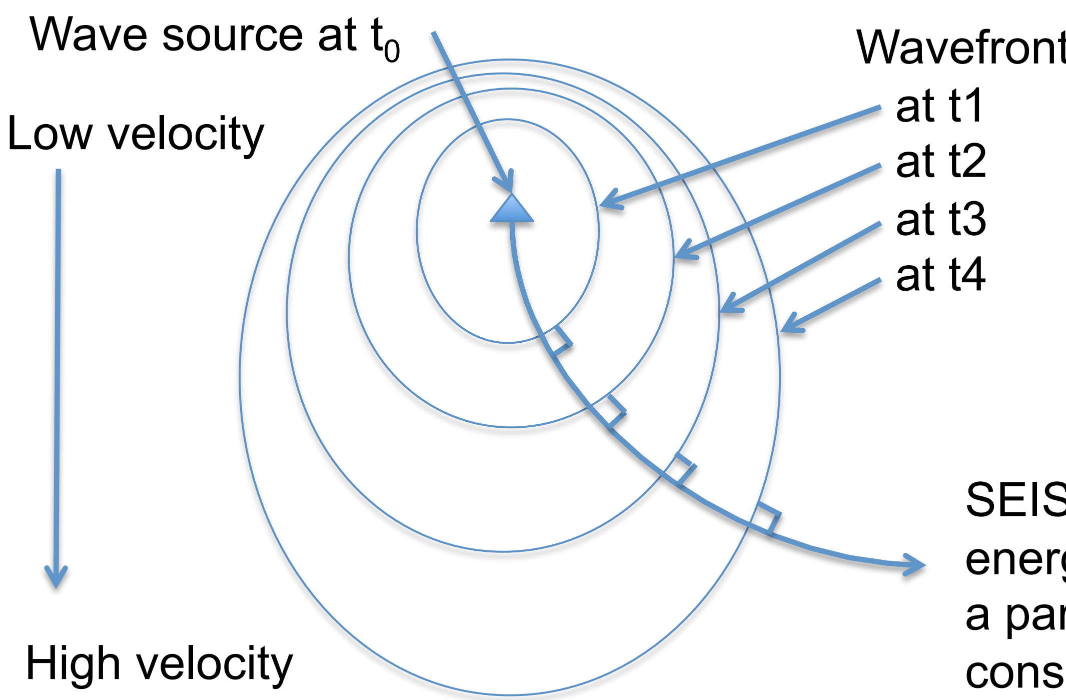
μ – rigidity

ρ – density

For a homogeneous medium, with uniform elastic properties, the wavefronts will be spherical, centered on the source.

Waves traveling through a medium that varies in material properties get distorted because the seismic velocity is not uniform. It is usually difficult to keep track of the *wavefront*, so we usually depict the wave propagation using *seismic rays*.

Seismic rays are the locus of points connecting vectors perpendicular to the wavefront as it varies in time, connecting material that has directly interacted along a specific path:

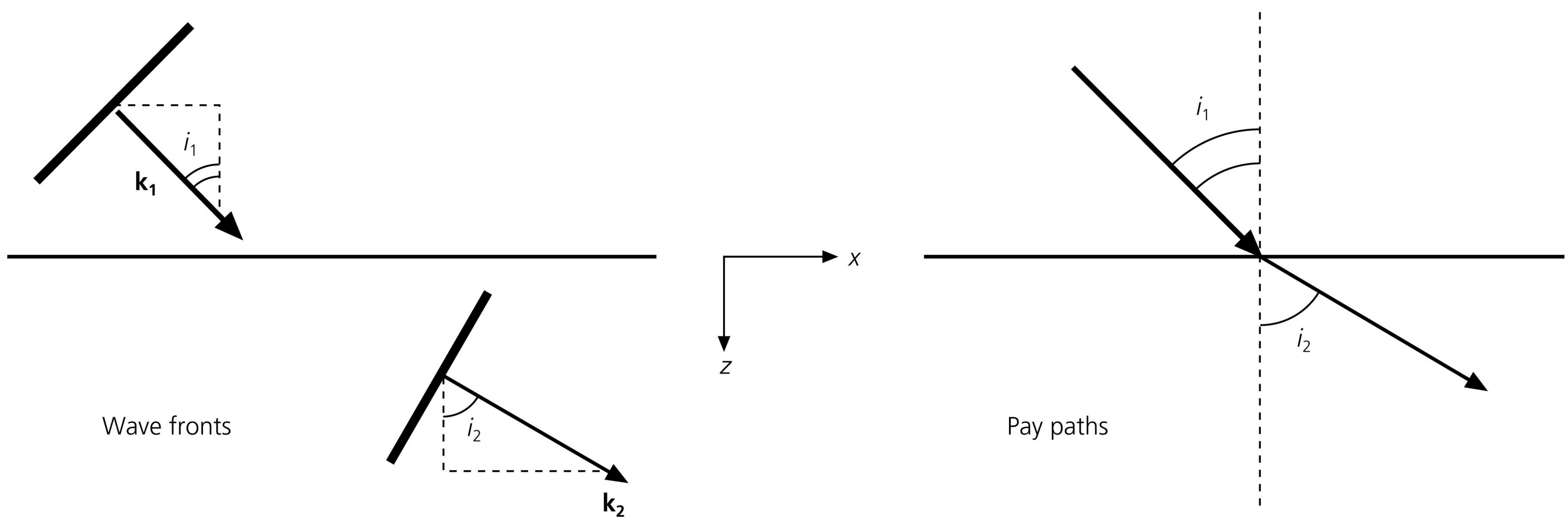


WAVEFRONT: Spatially distributed surface of the motions with the same value of (**x**-ct) (same phase) at a given instant of time. Spherical in constant velocity medium, distorted in variable velocity medium.

SEISMIC RAY – path followed by wave energy that left the source heading in a particular direction. Straight lines in constant velocity medium, curves in variable velocity medium.

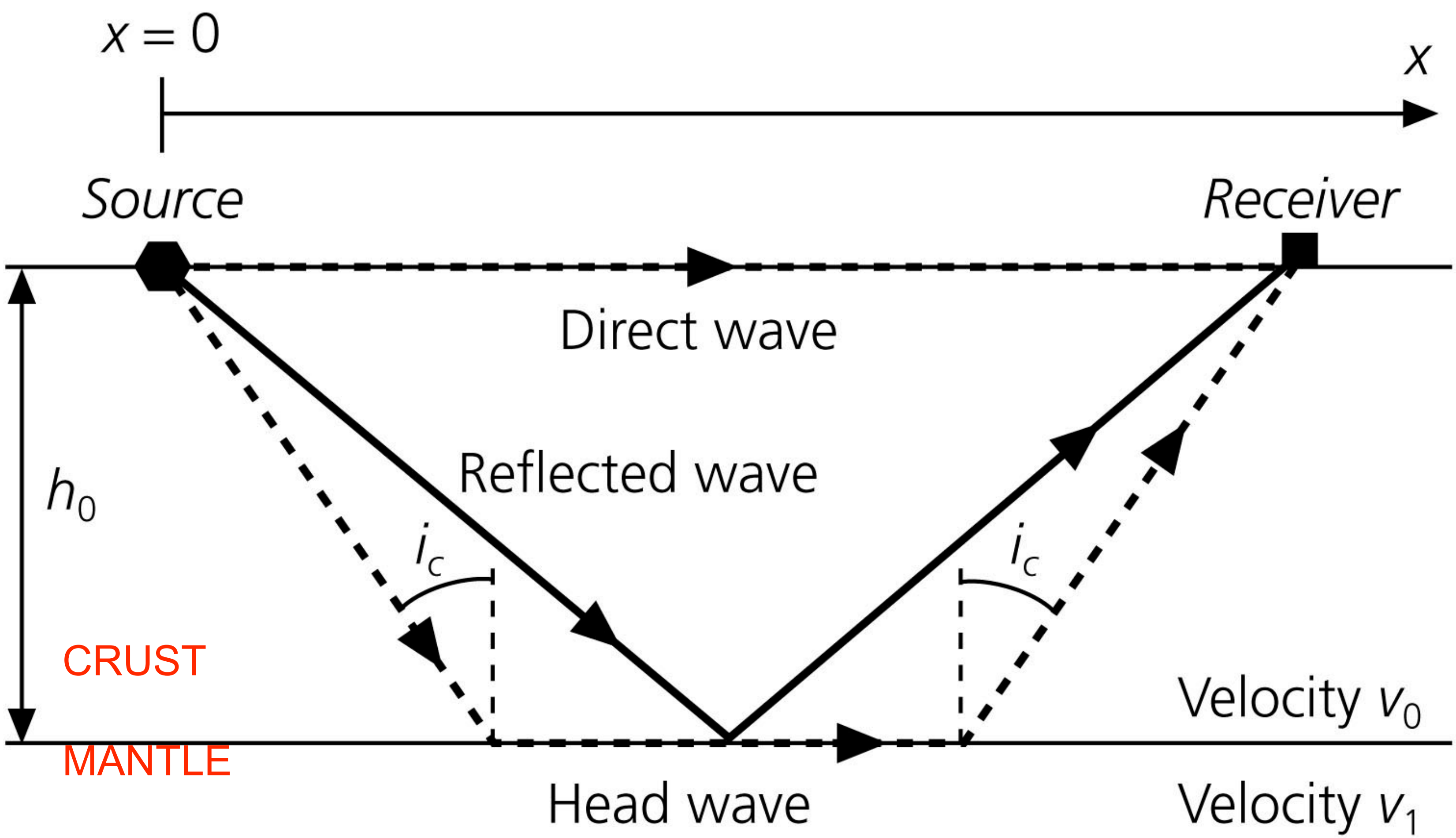
TRAVELLING BETWEEN MEDIA OF DIFFERENT VELOCITIES, WAVES BEND AND CHANGE AMPLITUDE

Figure 2.5-4: Change in wave front and direction during refraction.



Snell's Law for flat Layered structure gives a seismic ray parameter p = sin i / v. For a specific ray, the ray parameter is constant along the entire ray path.

Figure 3.2-1: Ray paths for a layer over a halfspace.



$V_1 > V_0$

Direct wave: $T_D(x) = x/v_0$

Reflected wave: $T_R(x) = 2(x^2/4 + h_0^2)^{1/2}/v_0$ (This curve is a hyperbola: $T_R^2(x) = x^2/v_0^2 + 4h_0^2/v_0^2$)

USE TRAVEL TIME CURVES TO FIND VELOCITY STRUCTURE VERSUS DEPTH

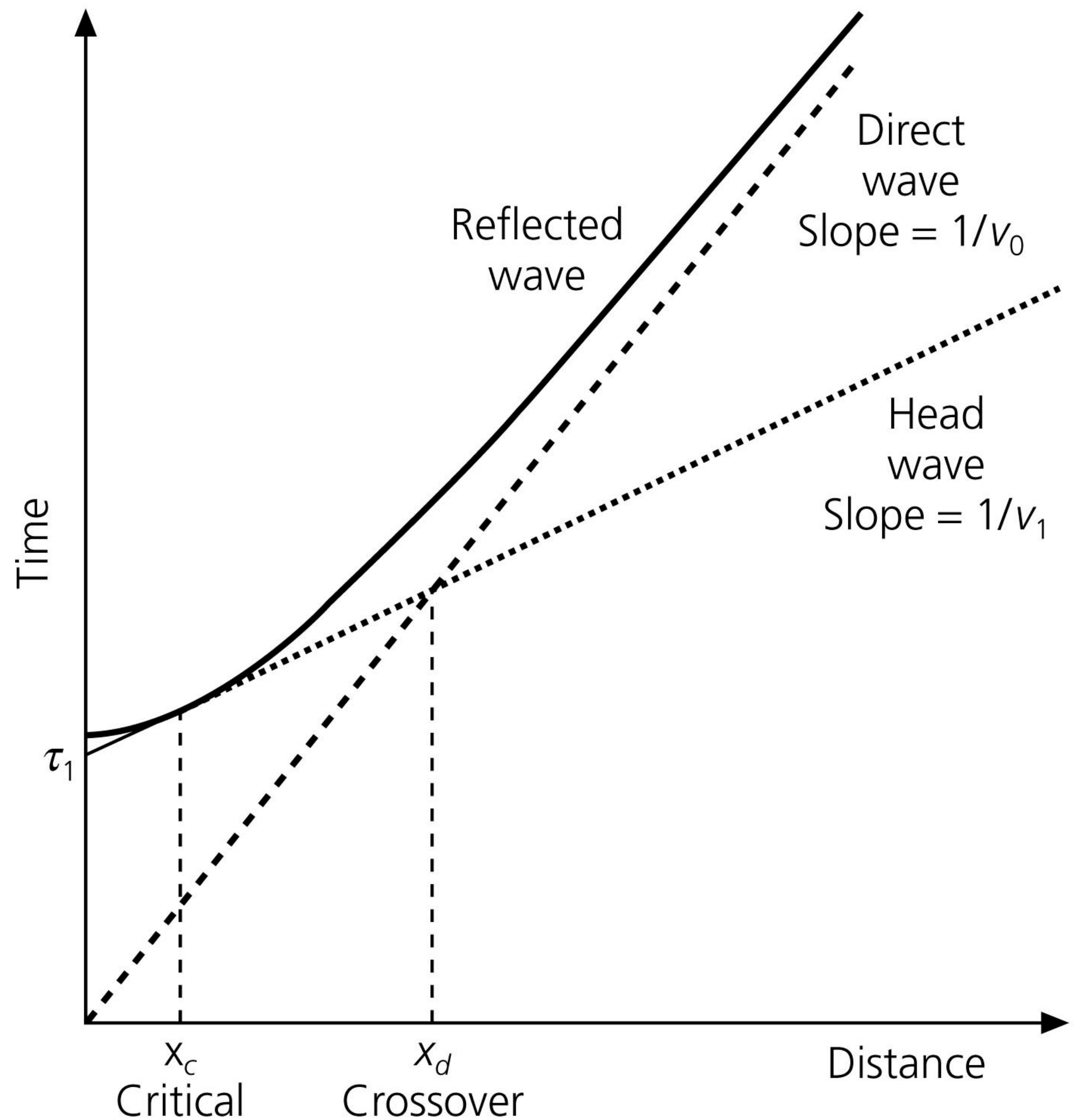
Direct wave:

$$T_D(x) = x/v_0$$

Reflected wave:

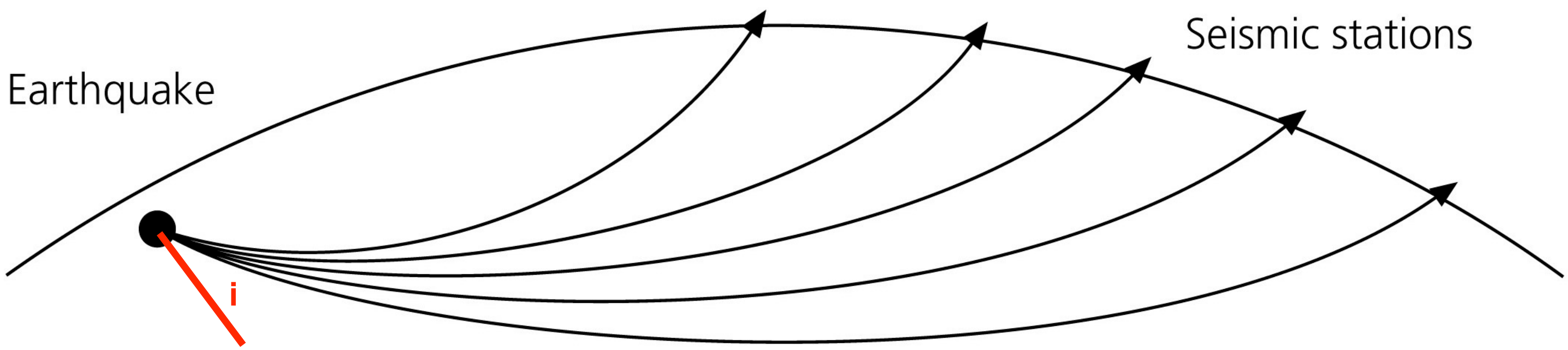
$$T_R(x) = 2(x^2/4 + h_0^2)^{1/2}/v_0$$

Figure 3.2-2: Travel time curve for rays in a layer over a halfspace.



SEISMIC RAY PATHS BEND AS VELOCITY INCREASES SMOOTHLY WITH DEPTH

Figure 1.1-2: Schematic ray paths for an increase in seismic velocity with depth.

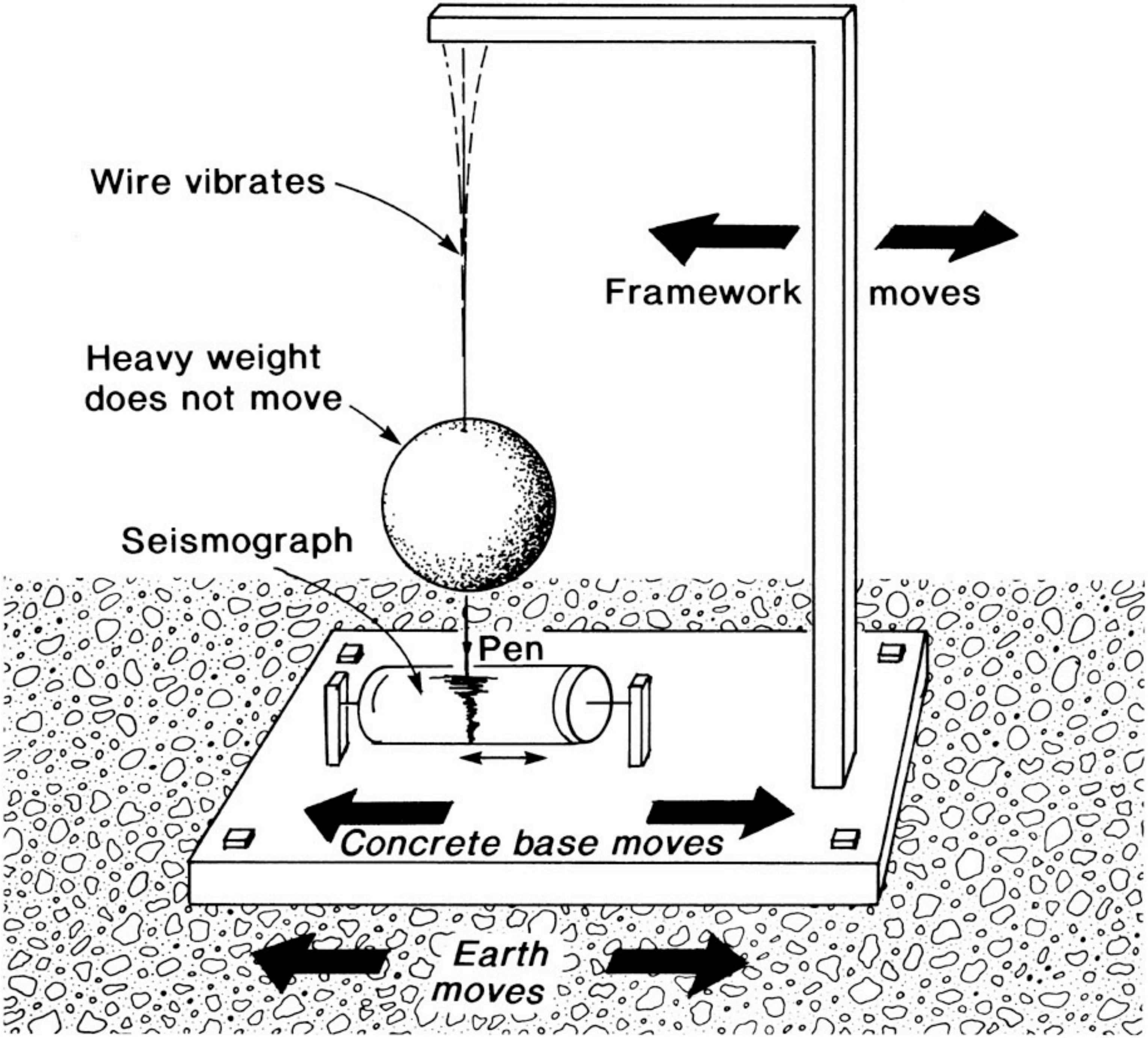


Snell's law for spherical earth with velocity v at radius r

Ray turns to incidence angle i so ray parameter p is constant along the ray.

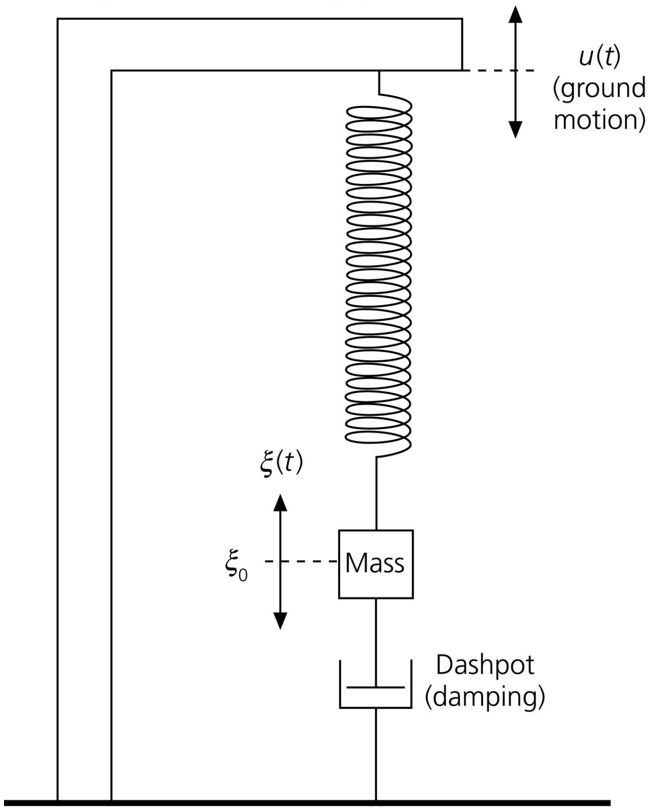
$$p = \frac{r \sin i}{v}$$

- So, to study the P and S waves, we need to record them. This requires measurements of ground motion at fixed positions as a function of time.
- Must measure full 3-D ground motion (up-down, north-south, east-west)
- Requires good clocks, and some measure of motion relative to a moving reference frame (not solved until 1870s).


Wire vibrates
Framework
moves
Heavy weight
does not move
Seismograph
Pen
Concrete base moves
Earth
moves

SEISMOMETER: MEASURE MOTION OF THE GROUND WITH INSTRUMENT ON THE GROUND

6.6-1: Diagram of a vertical seismograph.



This mechanical seismometer system is a damped simple harmonic oscillator. If the spring equilibrium length in the absence of ground motion is ξ_0, the spring exerts a force proportional to its extension from equilibrium as a function of time, $\xi(t) - \xi_0$, times a spring constant k. The dashpot, with damping constant d, exerts a force proportional to the velocity between the mass (m) and the earth. So, for a ground motion $u(t)$,

$$m \frac{d^2}{dt^2}[\xi(t) + u(t)] + d\frac{d\xi(t)}{dt} + \boxed{k[\xi(t) - \xi_0]} = 0. \quad (1)$$

If we define $\xi(t) - \xi_0$ as $\xi(t)$, the displacement relative to the equilibrium position, (1) becomes

$$m\ddot{\xi} + d\dot{\xi} + k\xi = -m\ddot{u} \quad (2)$$

or

$$\ddot{\xi} + 2\varepsilon\dot{\xi} + \omega_0^2\xi = -\ddot{u}, \quad (3)$$

where the single and double dots denote the first and second time derivatives, $\omega_0 = \sqrt{k/m}$ is the natural frequency of the undamped system, and the damping is described by $\varepsilon = d/(2m)$. This is a linear differential equation with constant coefficients

Pendulum seismograph consisting of a mass, spring, and dashpot

Figure 6.6-6: Coupling of the transducer of an electromagnetic seismograph to a galvanometer.

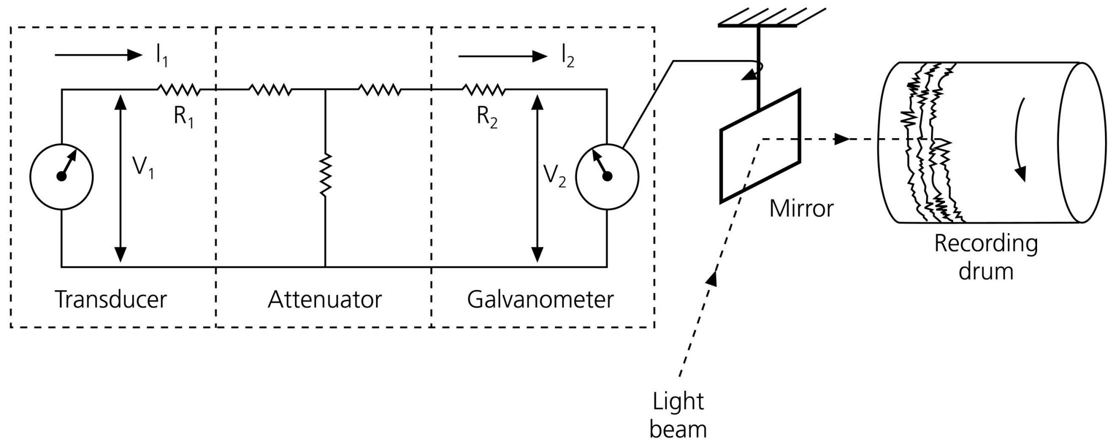


Figure 6.6-9: Sample WWSSN long-period vertical-component seismogram for one day.

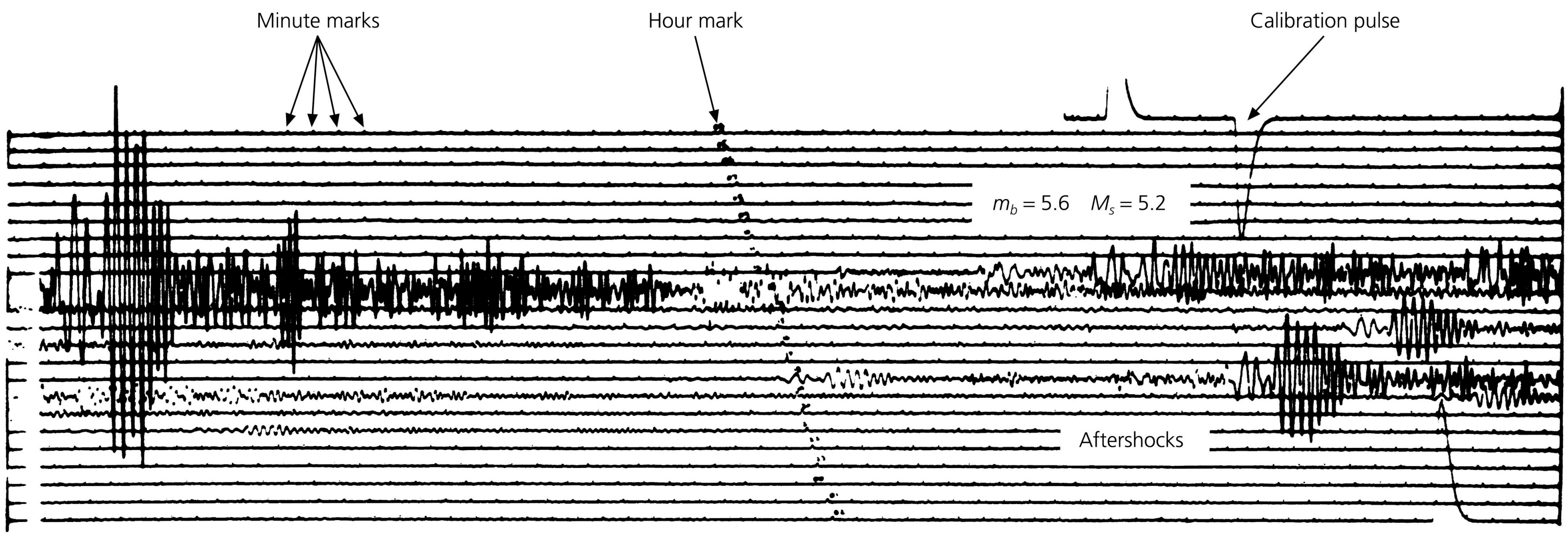
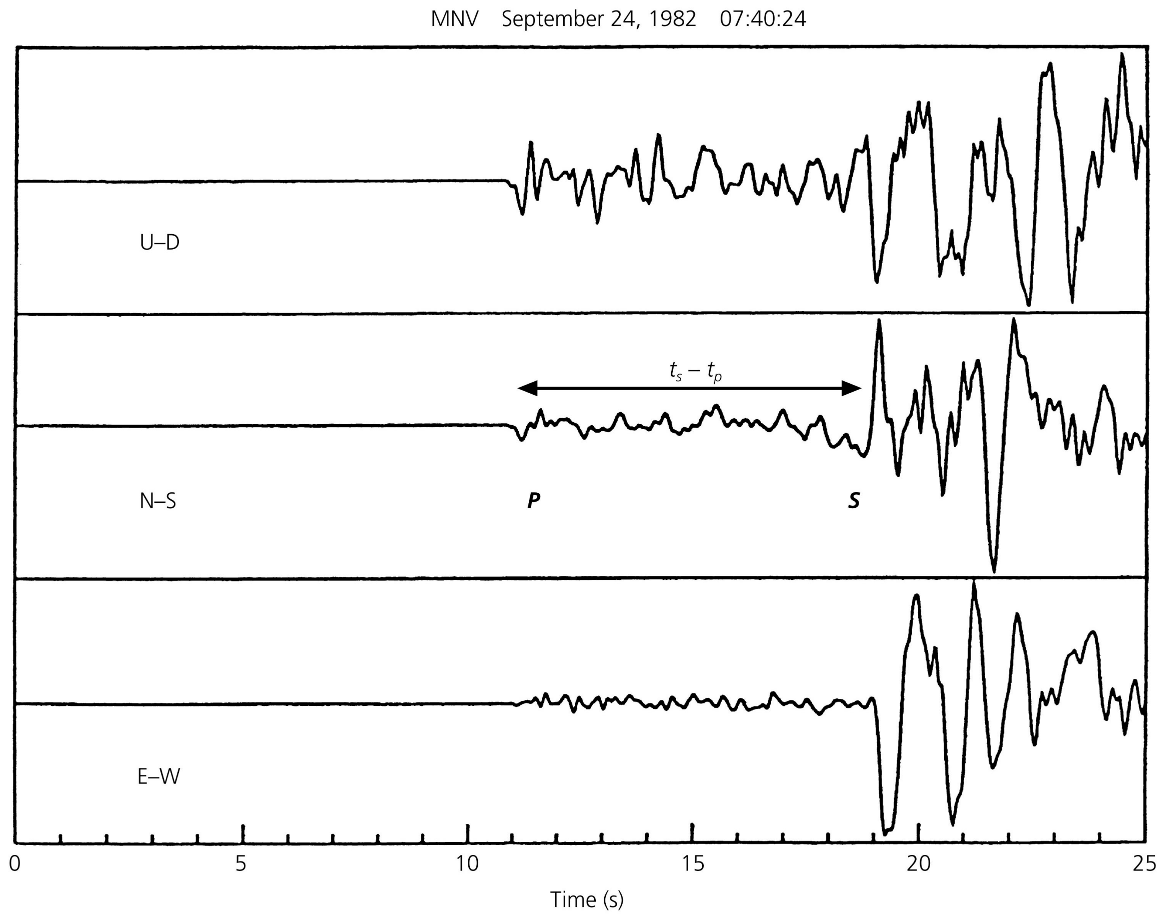


Figure 2.4-9: Seismograms recorded 64 km from a small, shallow earthquake.



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$$t_s - t_p = x(1.0/3.2 - 1.0/5.5) = x/7.6$$

Figure 3.5-3: Travel time data and curves for the IASP91 model.

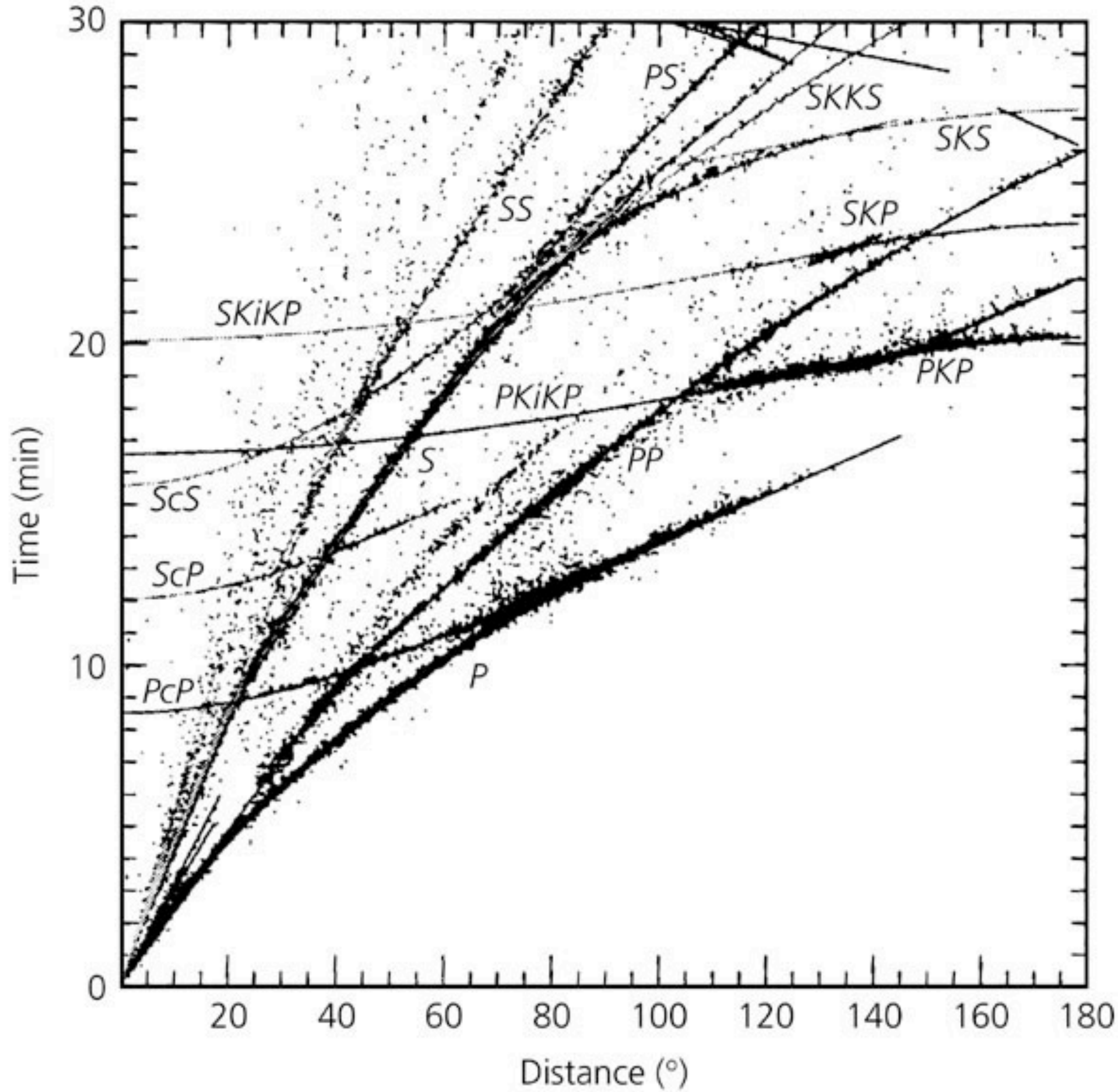
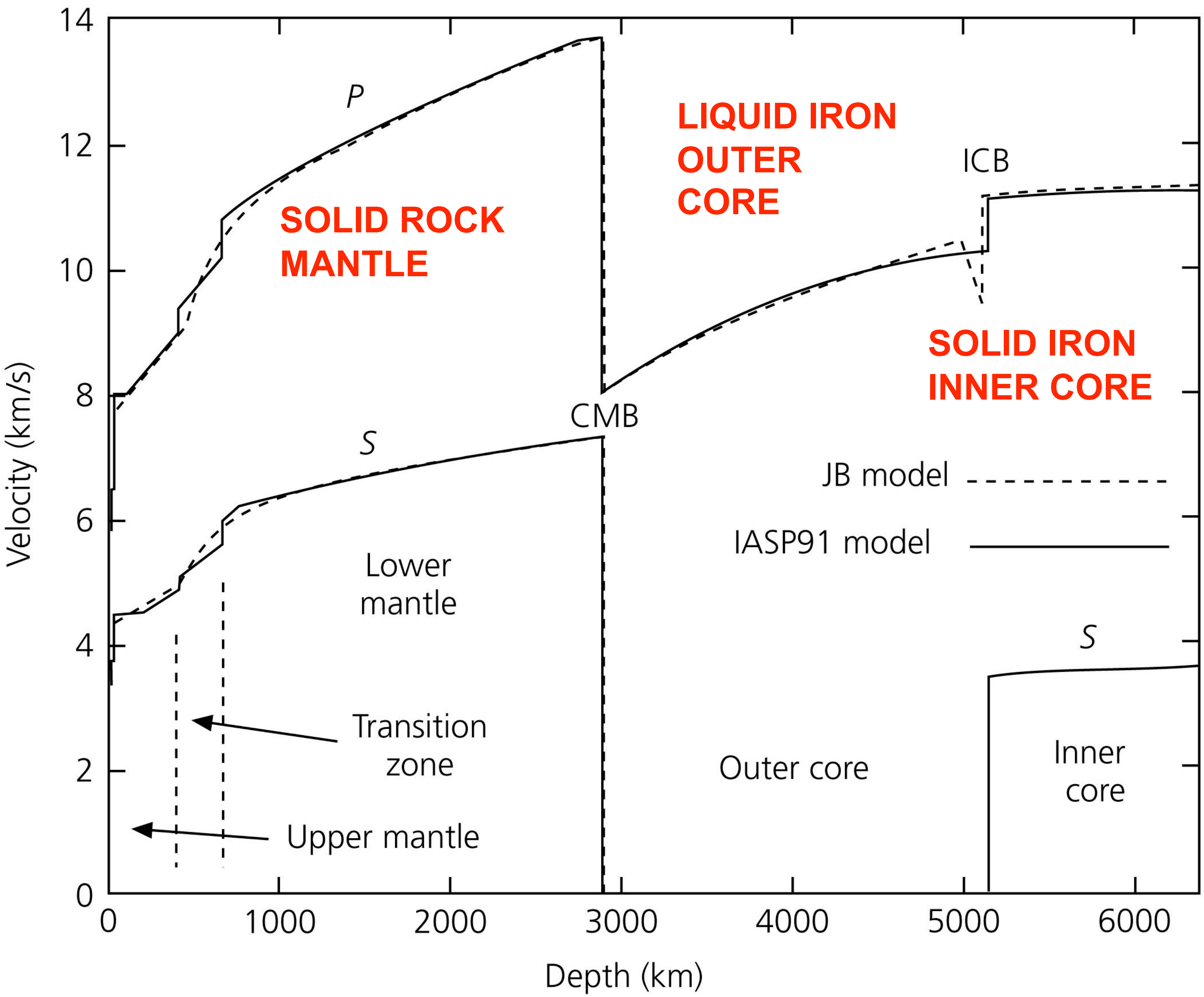
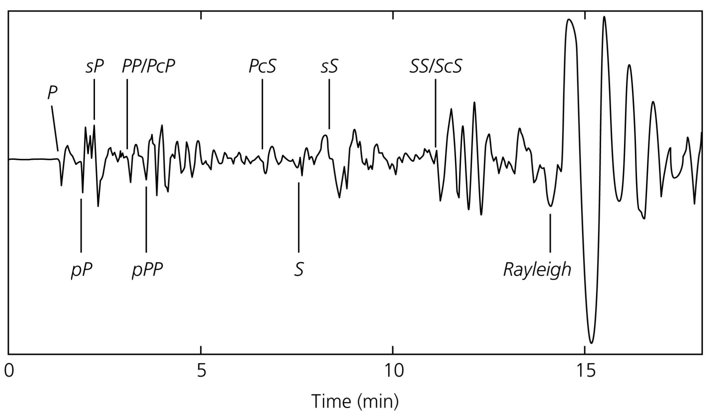


Figure 3.5-1: Comparison of the J-B and IASP91 earth models.



TRAVEL TIME CURVES

Figure 1.1-3: Example of seismogram,



showing accompanying ray paths.

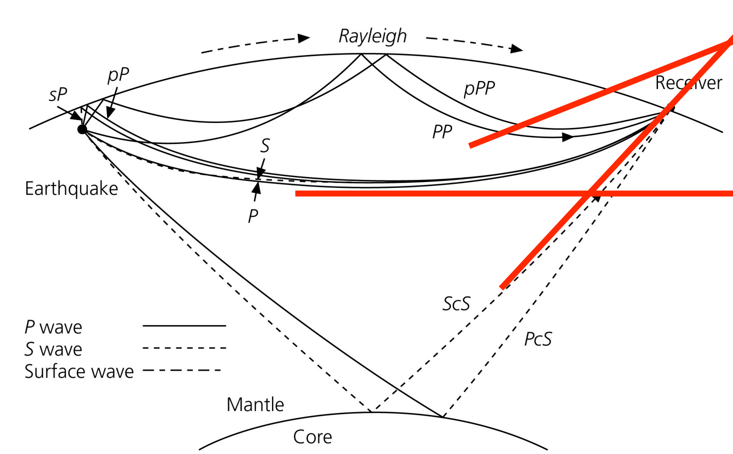
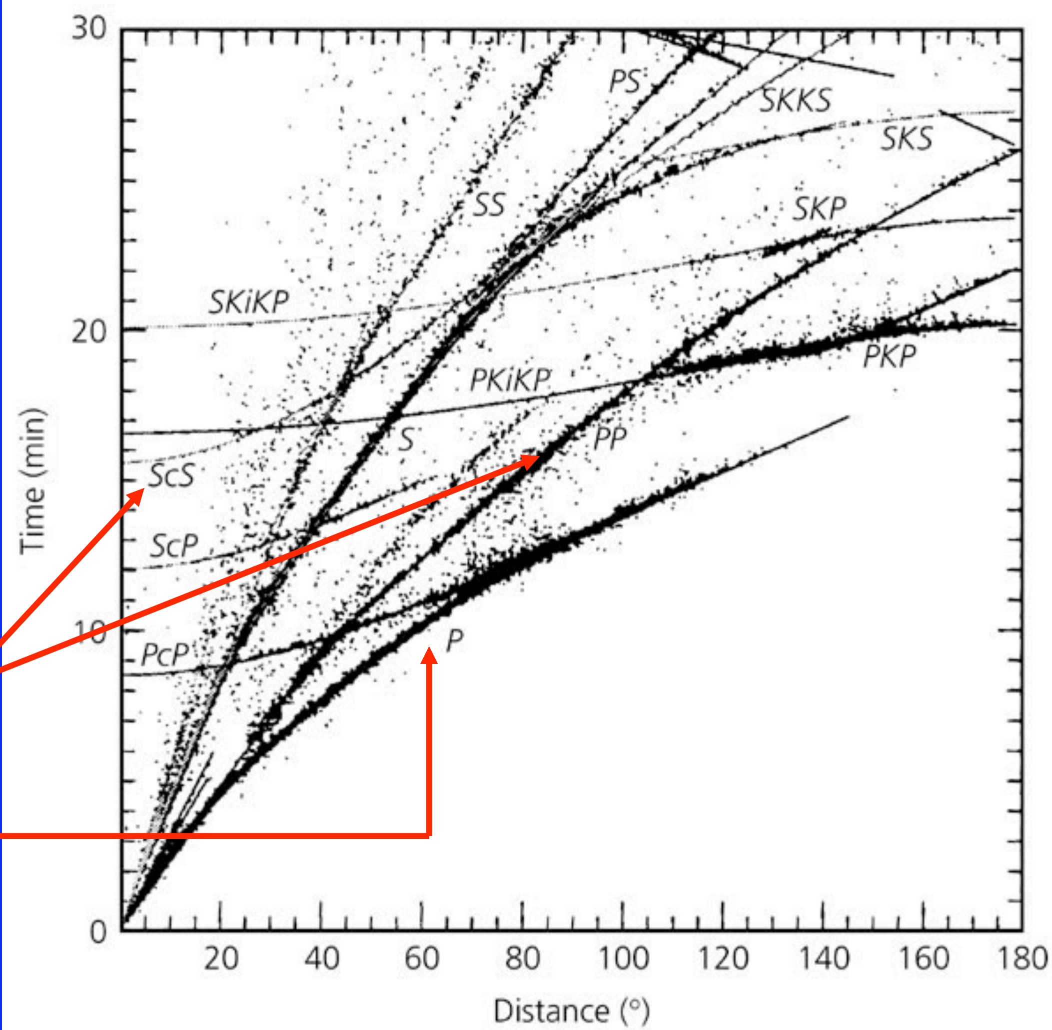
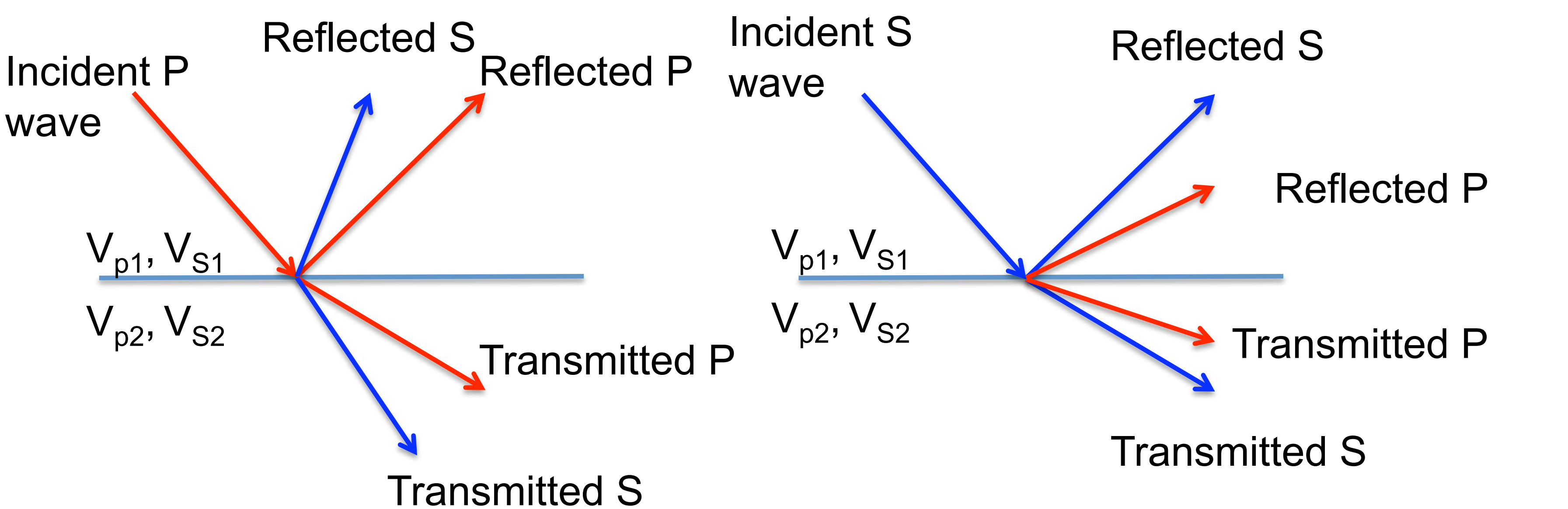


Figure 3.5-3: Travel time data and curves for the IASP91 model.



So, much of the complexity of observed seismograms is caused by the expanding P and S waves encountering the Earth's surface, where energy can reflect down, as well as the core, where energy can reflect off and transmit across. But additional complexity is produced by the property of seismic waves that is unique to elastic waves; a wave type (P or S) incident on an abrupt change in material properties (surface, rock layer interface, core-mantle boundary), can convert some of its energy into the other wave type (P energy can convert to S, or S energy to P). Thus, and incident P wave can produce reflected P, transmitted P, reflected S and transmitted S. How much energy goes into each phase is determined by solving for continuity of stress and displacement at the boundary – this gives *reflection* and *transmission coefficients*.



Earth’s interior structure and seismic raypaths that are used to determine the Earth structure.

http://www.iris.edu/hq/files/programs/education_and_outreach/lessons_and_resources/images/ExplorEarthPoster.jpg

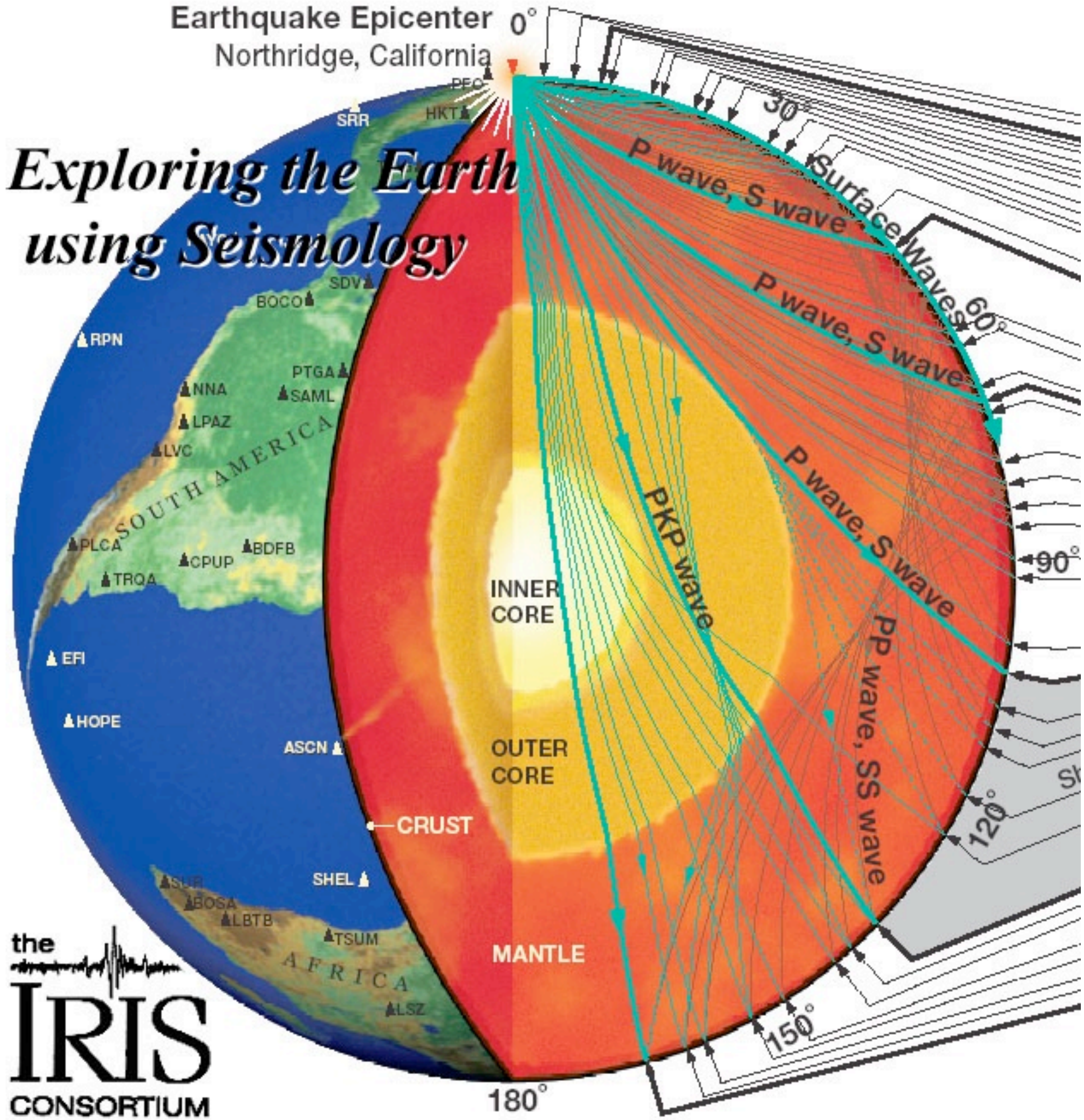
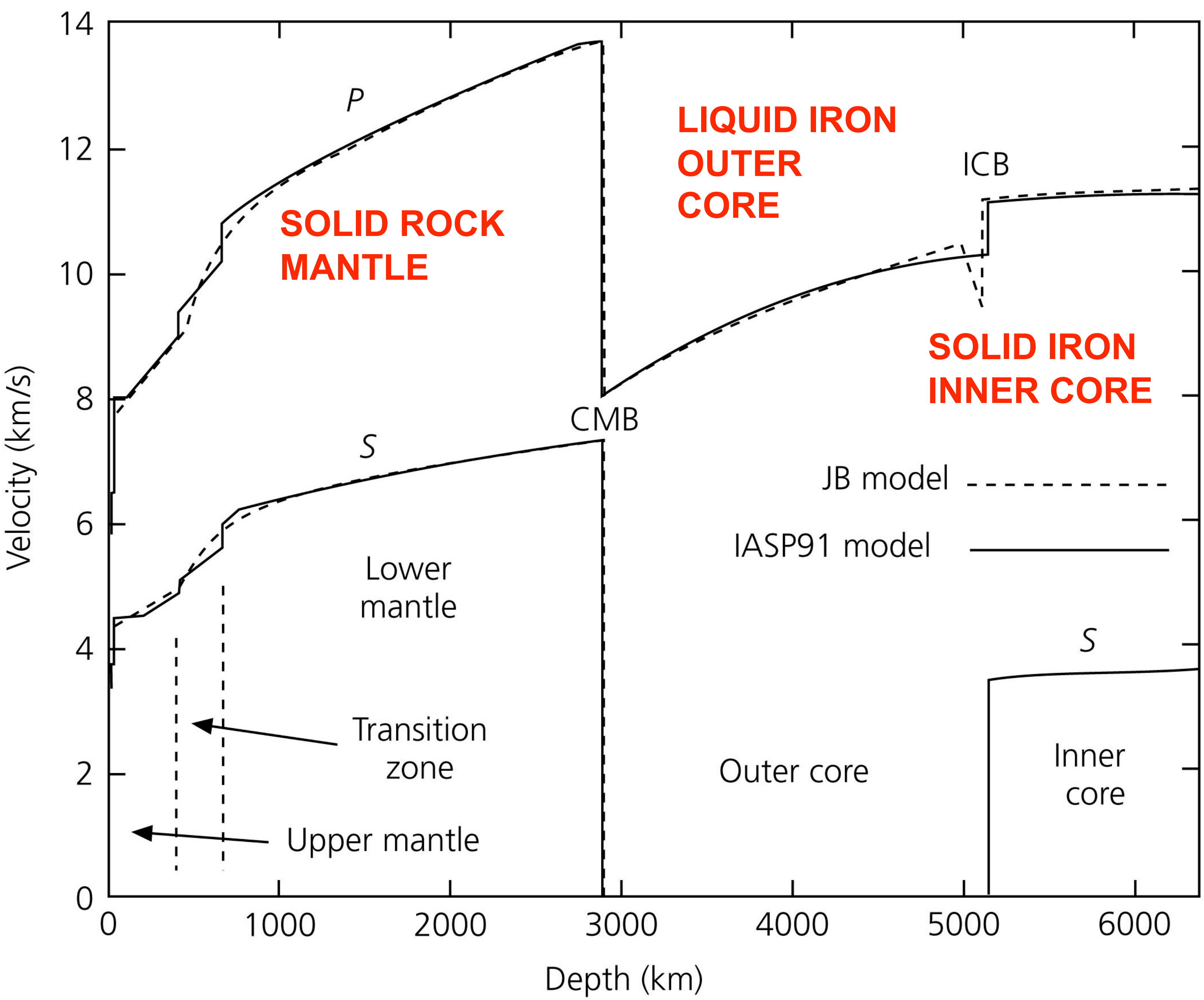
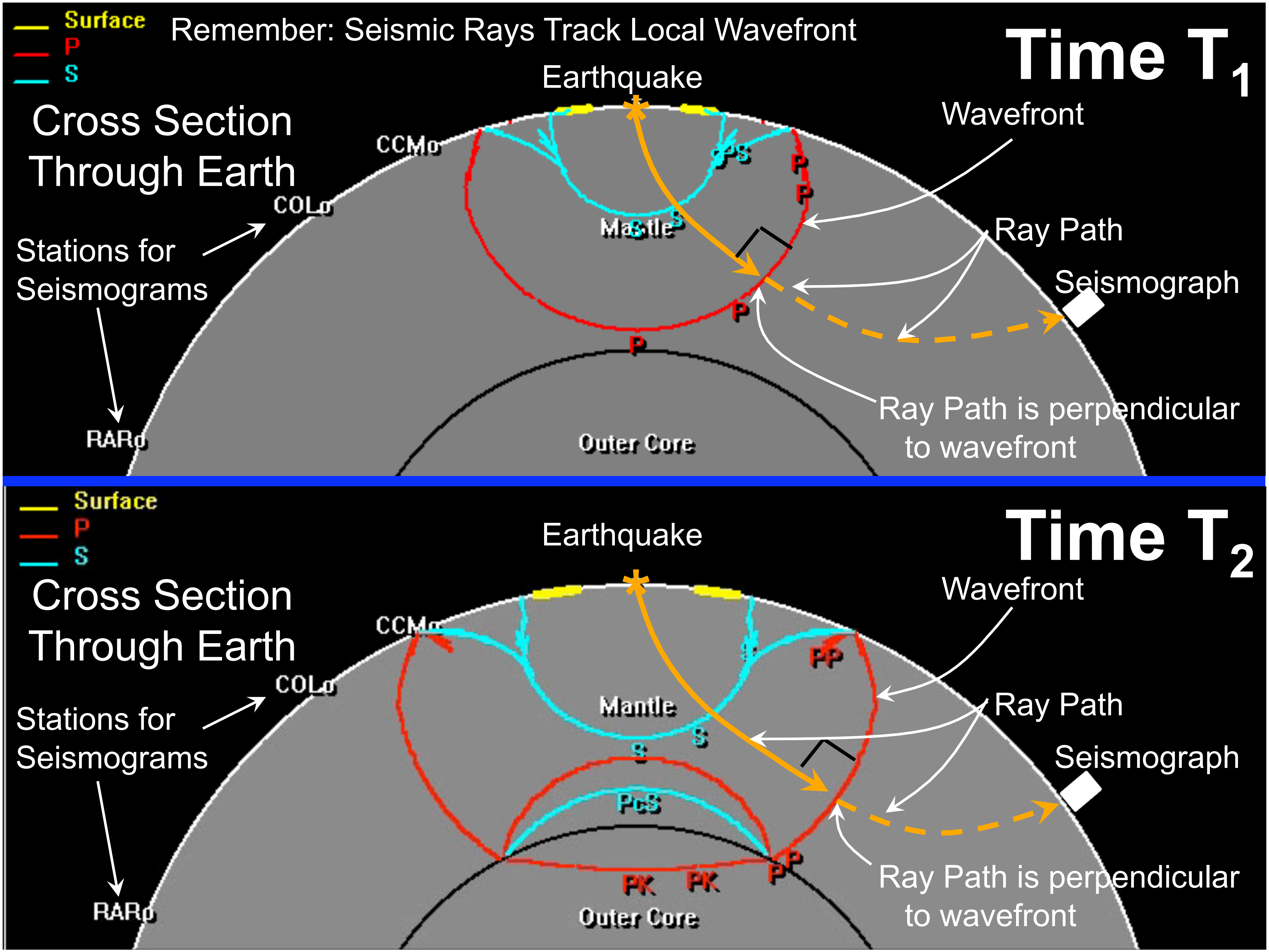


Figure 3.5-1: Comparison of the J-B and IASP91 earth models.





Surface
P
S
Remember: Seismic Rays Track Local Wavefront
Time T_1
Earthquake
Cross Section Through Earth
Wavefront
CCMo
COLo
Mantle
Ray Path
Stations for Seismograms
Seismograph
RARo
Outer Core
Ray Path is perpendicular to wavefront
Time T_2
PP
PcS
PK

Computed motions in A 3D spherical Velocity structure show the S wavefront as It folds and reflects off of the surface and the core-mantle boundary.

Figure 3.5-19: Snapshots of a synthetic *SH* wave field at various times.

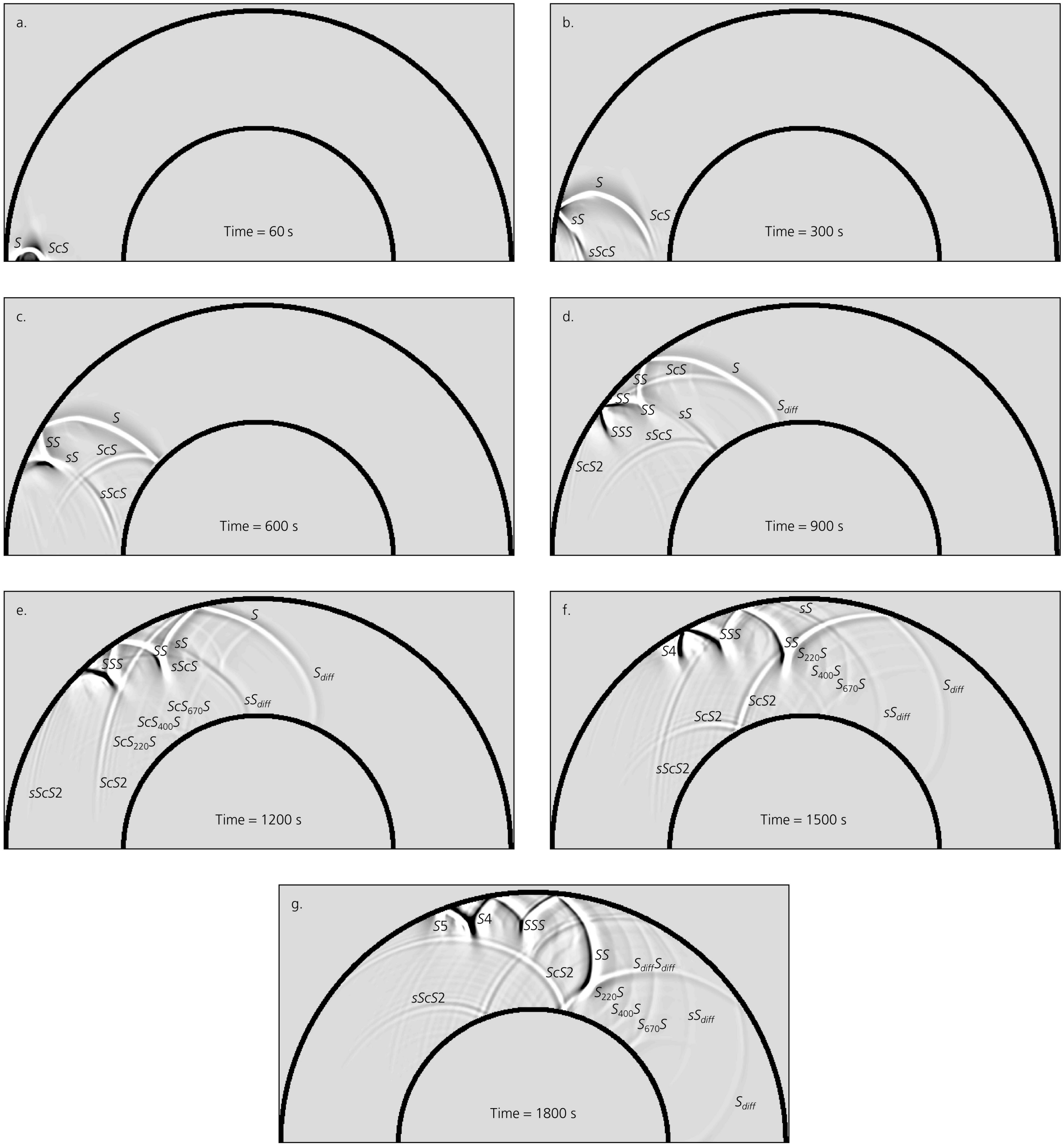
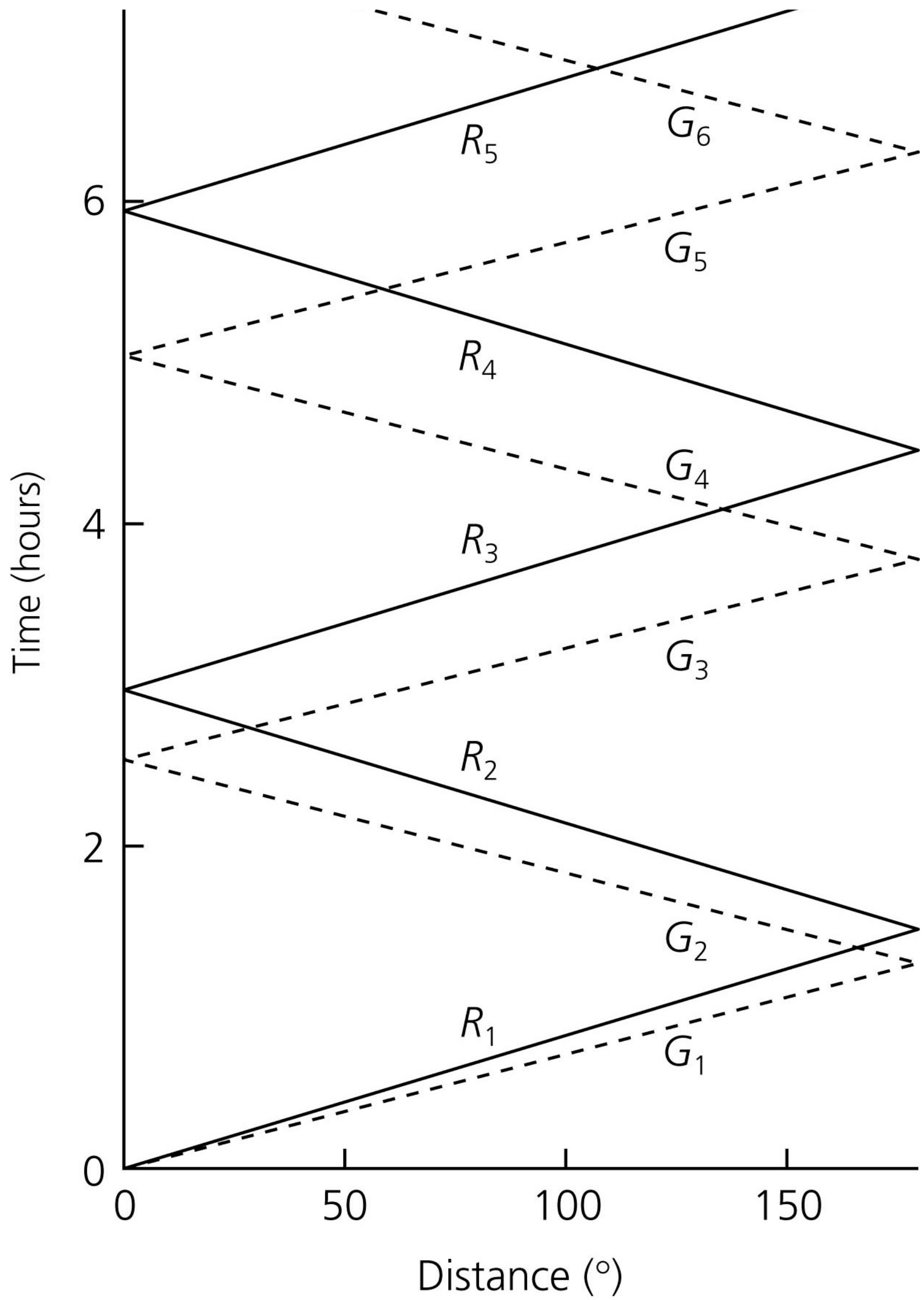
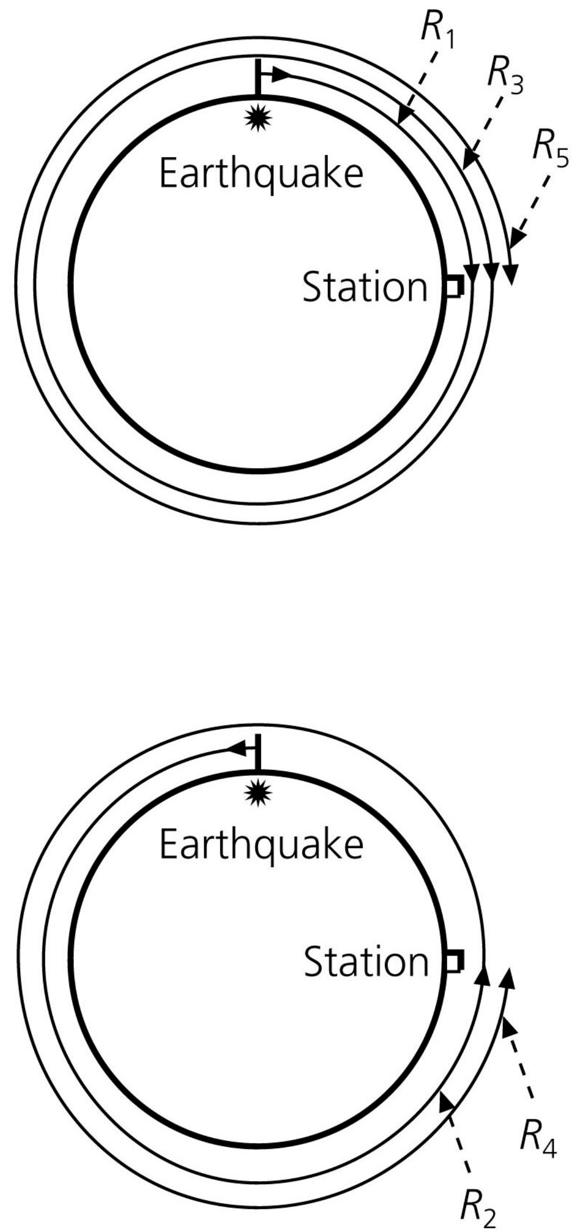


Figure 2.7-4: Six-hour stacked IDA record section.

Figure 2.7-3: Multiple surface waves circle the earth.





Waves from a single earthquake spread around the world multiple times. If the earthquake is big, like the 2004 Sumatra earthquake, the motions can be detected for hours and even days on sensitive seismometers.

The records here are from Global Seismic Network stations operated by the Incorporated Research Institution for Seismology (IRIS), in partnership with the NSF and USGS

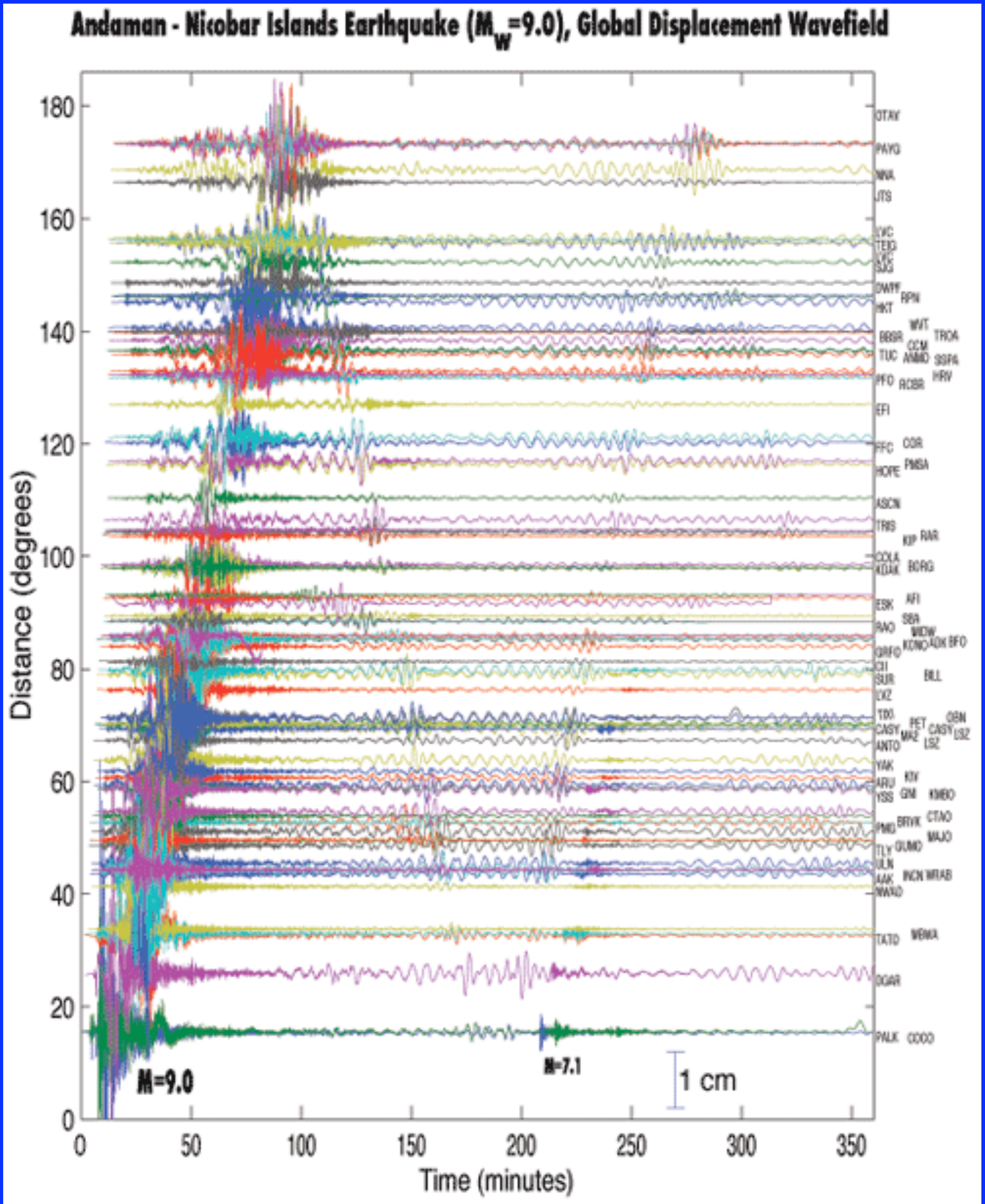
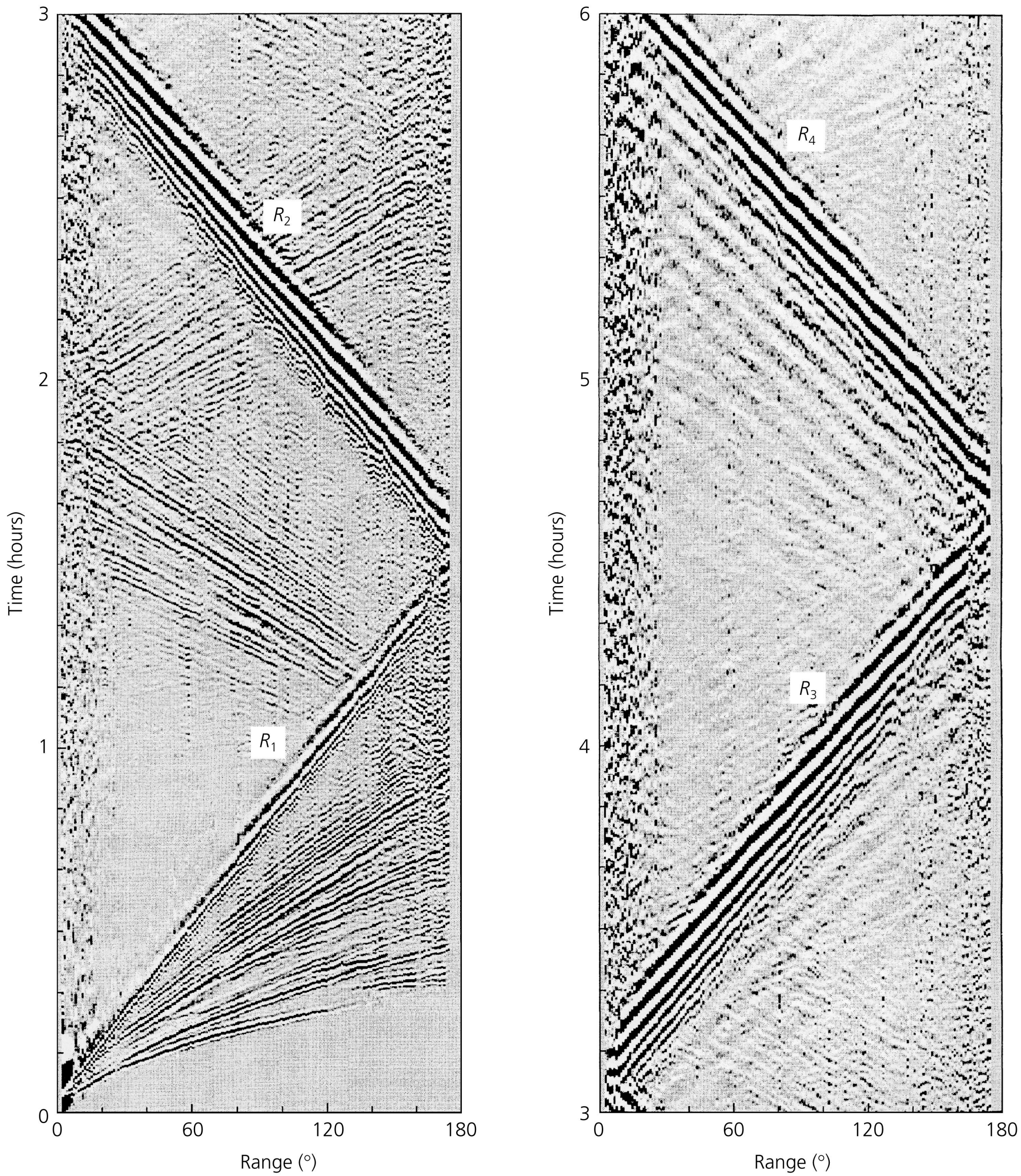
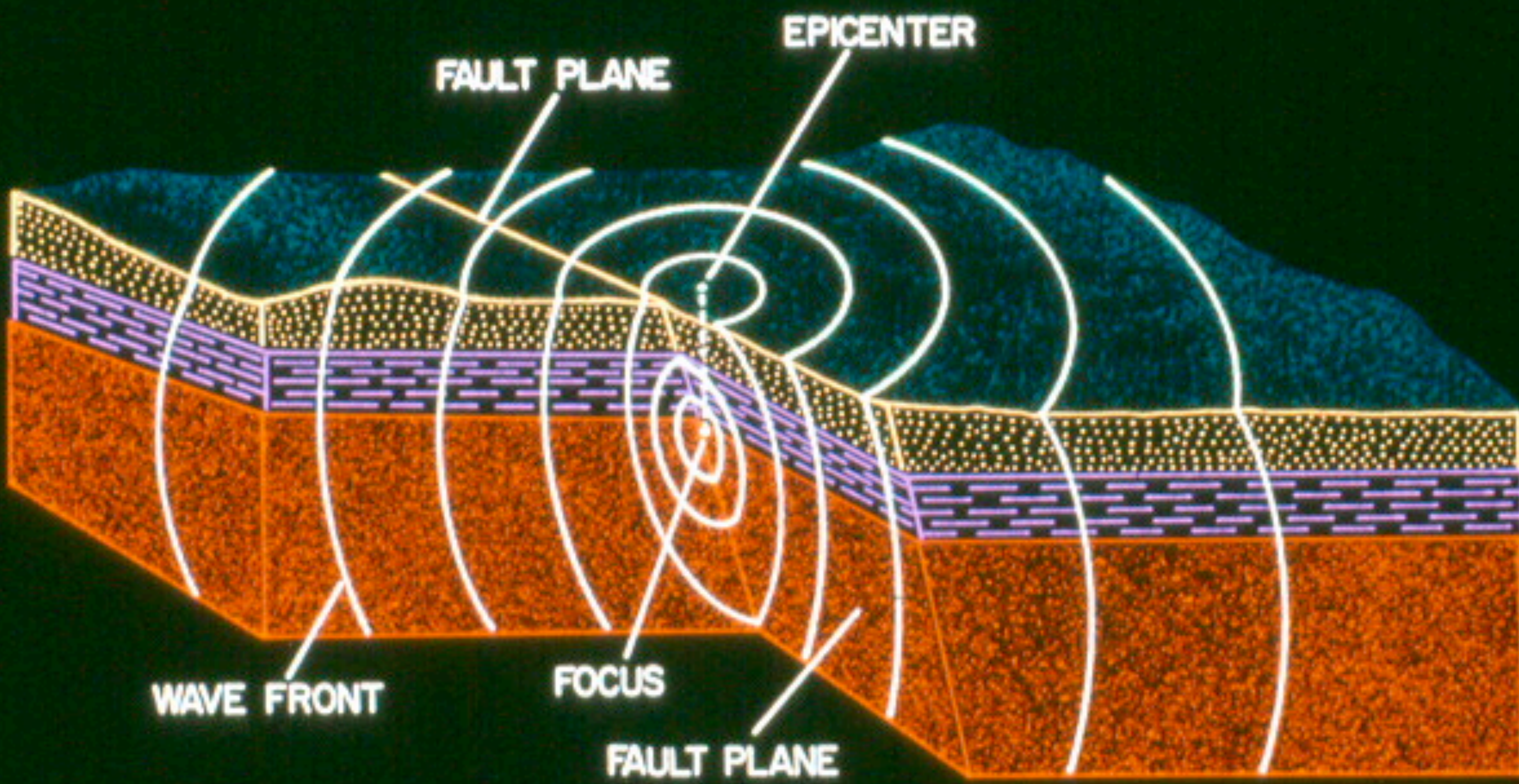
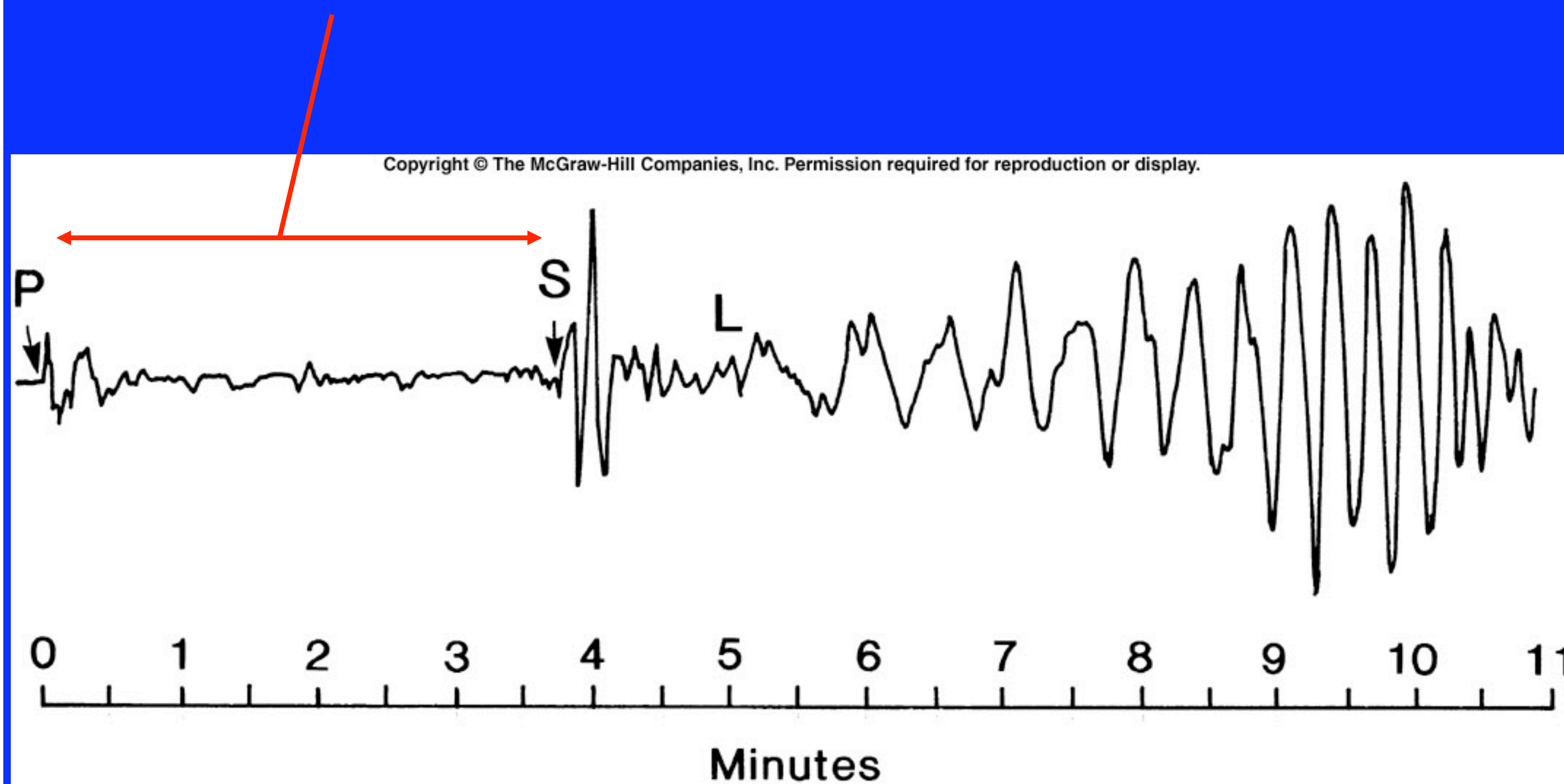


Figure 2.7-4: Six-hour stacked IDA record section.





EPICENTER AND FOCUS OF EARTHQUAKE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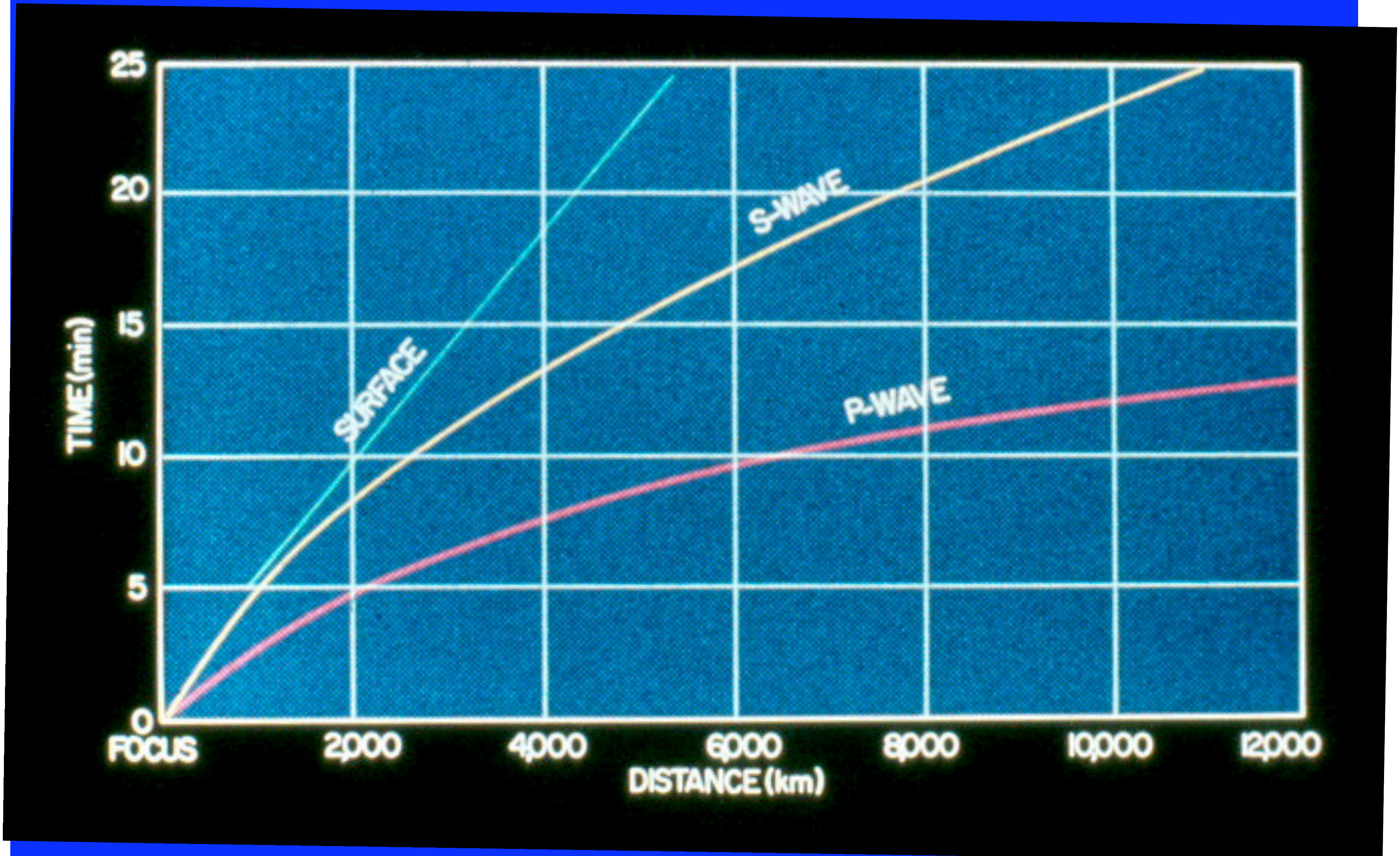
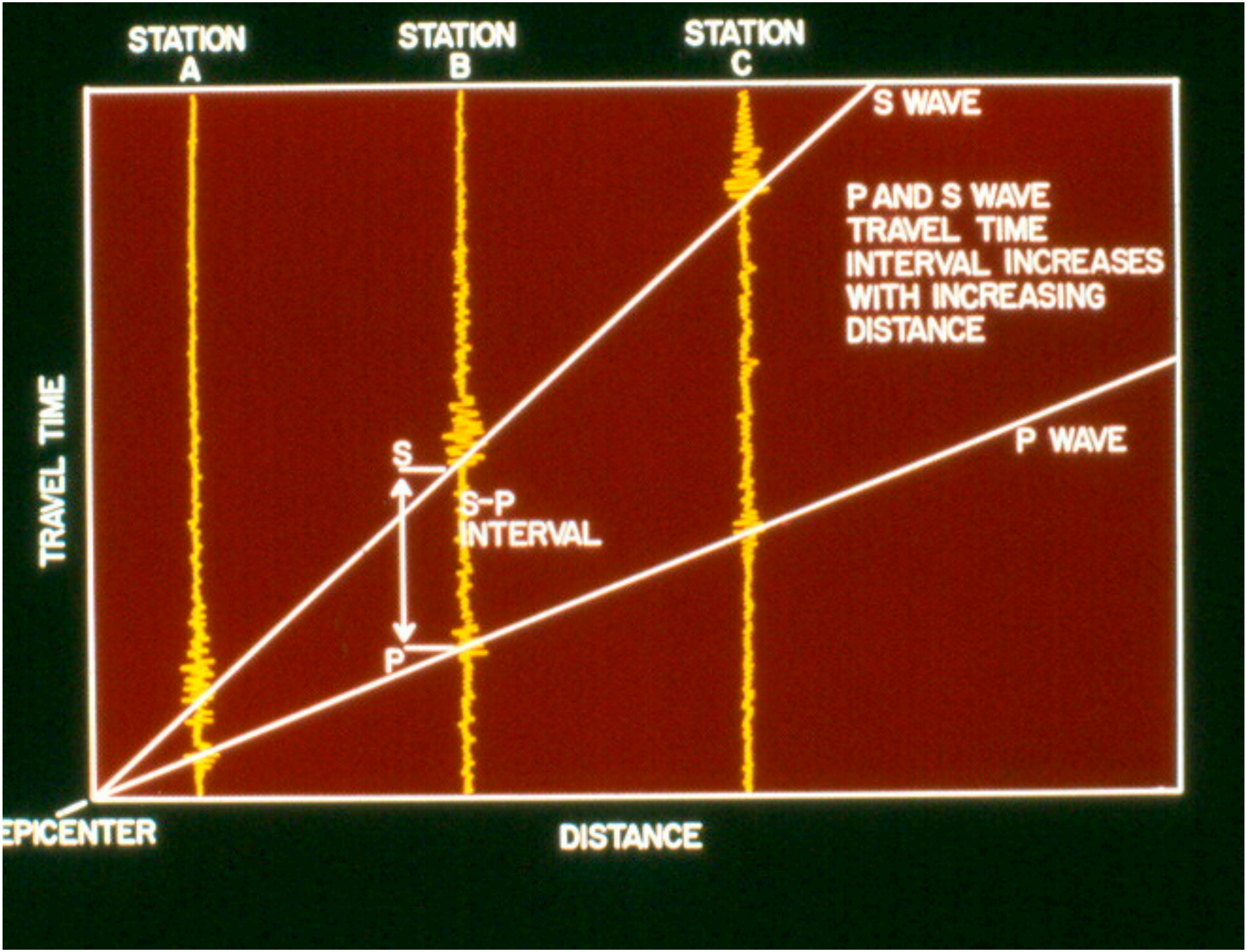
P
S
L
0
1
2
3
4
5
6
7
8
9
10
11
Minutes



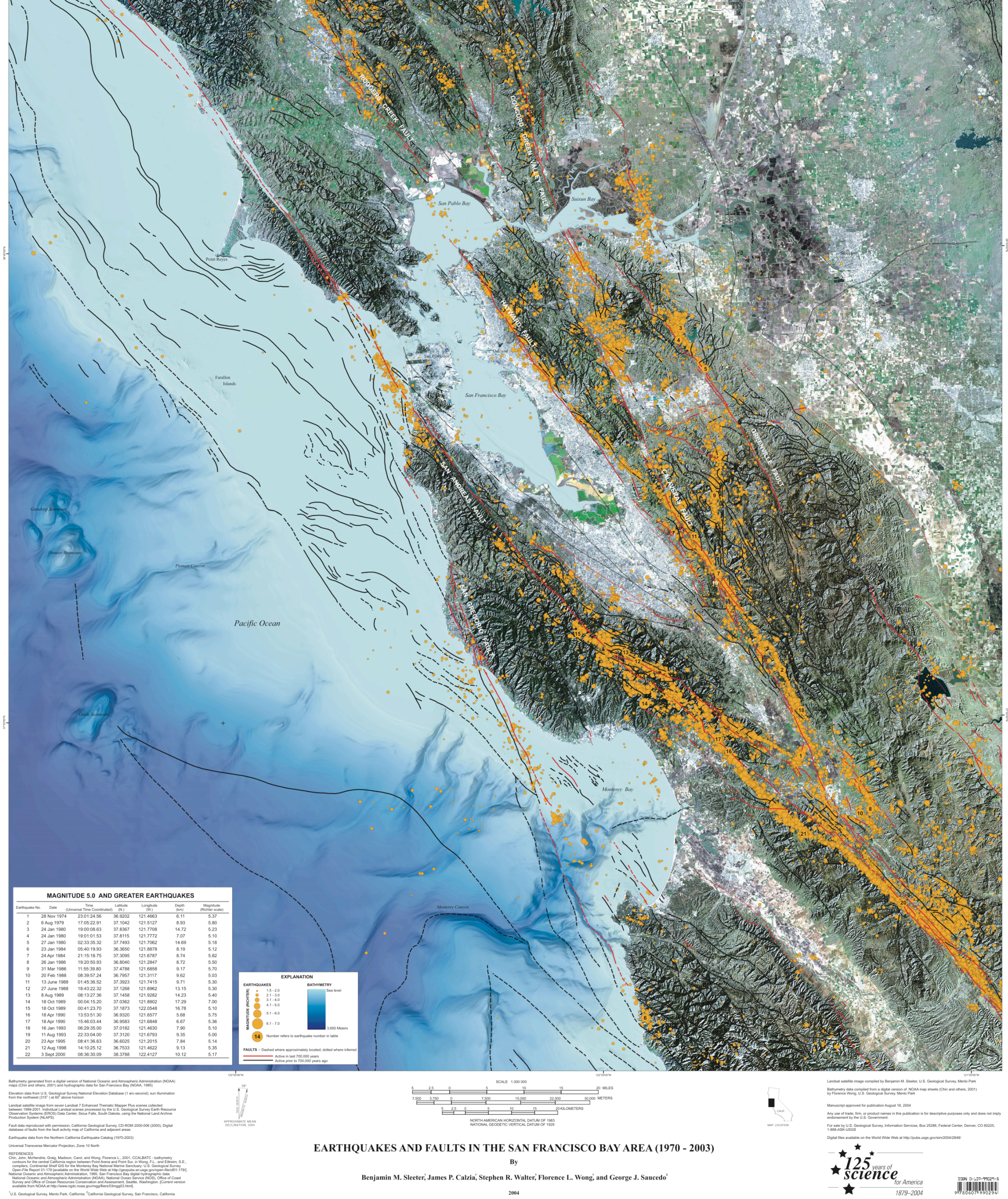
25
20
15
10
5
0
TIME (min)
SURFACE
S-WAVE
P-WAVE
FOCUS
2,000
4,000
6,000
8,000
10,000
12,000
DISTANCE (km)


STATION A
STATION B
STATION C
S WAVE
P AND S WAVE TRAVEL TIME INTERVAL INCREASES WITH INCREASING DISTANCE
P WAVE
S
S-P INTERVAL
P
TRAVEL TIME
EPICENTER
DISTANCE

We can measure S-P differential arrival times at several stations. From previous calibration of that time difference as a function of distance from a source (indicated by the regional travel time curve), we estimate the distance from each station to the Source and the origin time. Intersections of circles having radii of corresponding distances should 'triangulate' the unique position of the source at the correct distance from 3 or more stations (or just P arrivals from many stations using a non-linear least squares Inversion).

So we can routinely locate earthquakes:

This requires distribution of seismometers with good clocks, computer or analysts to pick arrivals, and solution of location problem.



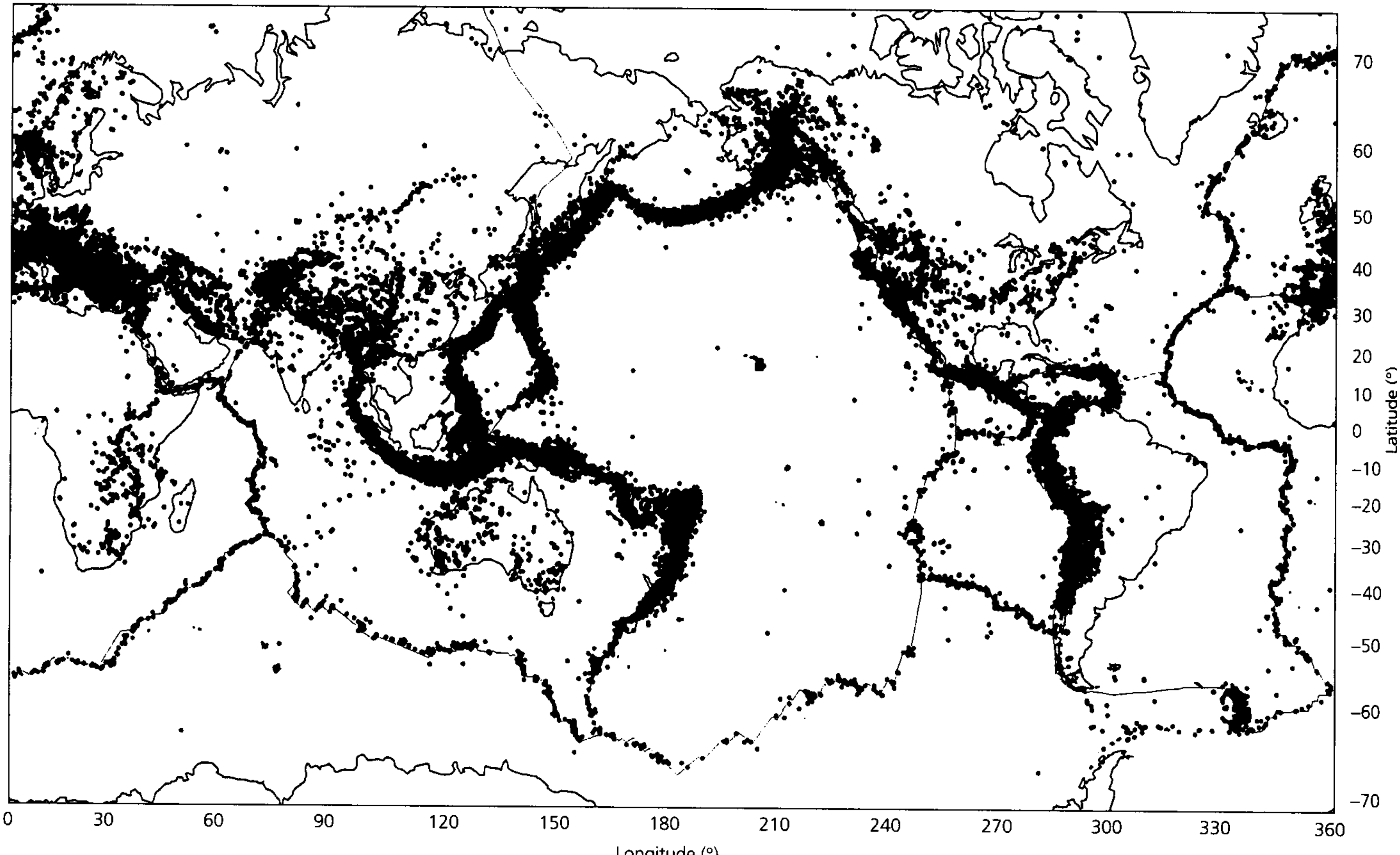
MAGNITUDE 5.0 AND GREATER EARTHQUAKES

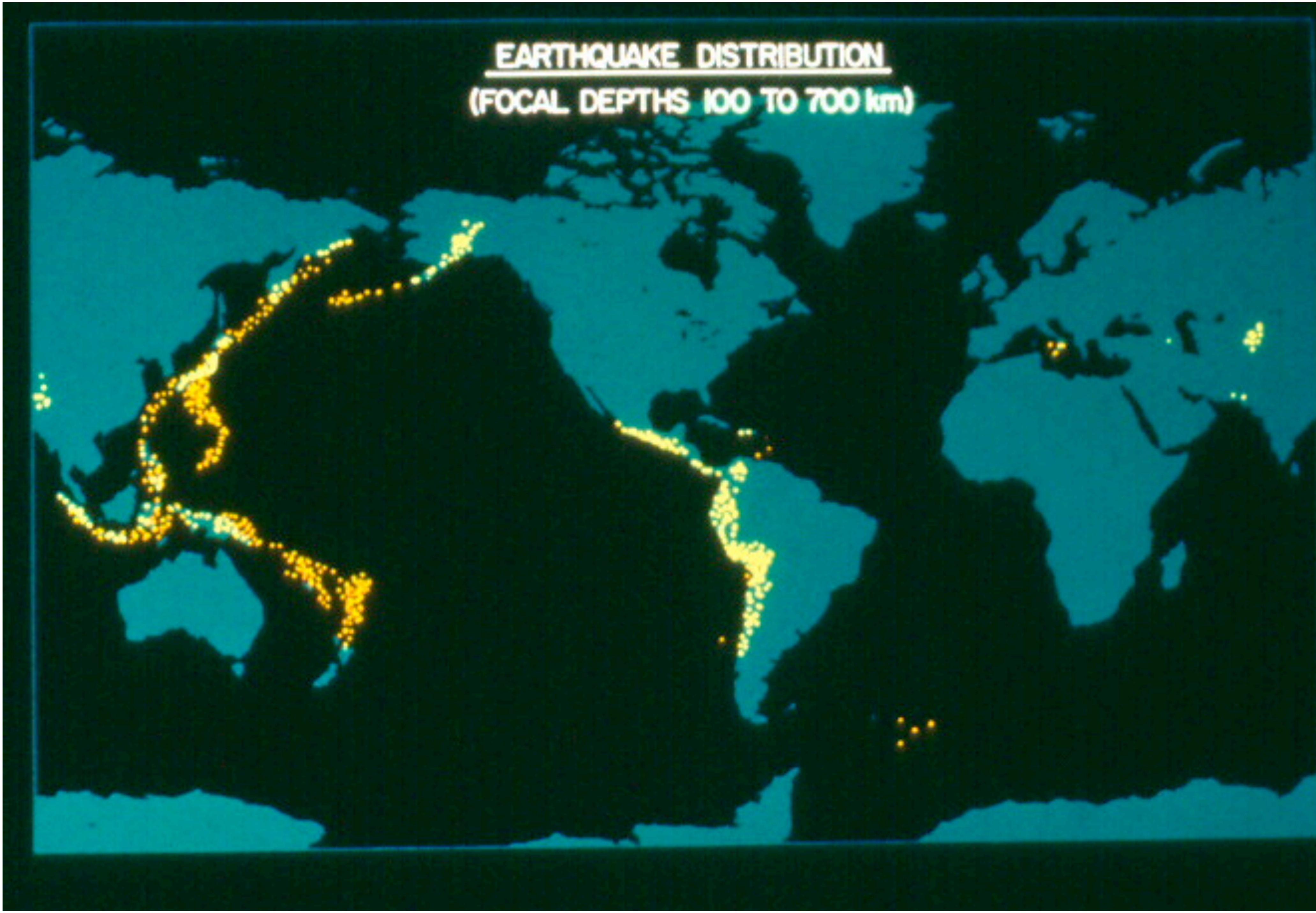
Earthquake No.	Date	Time (Universal Time Coordinated)	Latitude (N.)	Longitude (W.)	Depth (km)	Magnitude (Richter scale)
1	28 Nov 1974	23:01:24.56	36.9202	121.4663	6.11	5.37
2	6 Aug 1979	17:05:22.91	37.1042	121.5127	8.93	5.80
3	24 Jan 1980	19:00:08.63	37.8367	121.7708	14.72	5.23
4	24 Jan 1980	19:01:01.53	37.8115	121.7772	7.07	5.10
5	27 Jan 1980	02:33:35.32	37.7493	121.7062	14.69	5.18
6	23 Jan 1984	05:40:19.93	36.3650	121.8878	8.19	5.12
7	24 Apr 1984	21:15:18.75	37.3095	121.6787	8.74	5.82
8	26 Jan 1986	19:20:50.93	36.8040	121.2847	8.72	5.50
9	31 Mar 1986	11:55:39.80	37.4788	121.6858	9.17	5.70
10	20 Feb 1988	08:39:57.24	36.7957	121.3117	9.62	5.03
11	13 June 1988	01:45:36.52	37.3923	121.7415	9.71	5.30
12	27 June 1988	18:43:22.32	37.1268	121.8962	13.15	5.30
13	8 Aug 1989	08:13:27.36	37.1458	121.9282	14.23	5.40
14	18 Oct 1989	00:04:15.20	37.0362	121.8802	17.29	7.00
15	18 Oct 1989	00:41:23.70	37.1873	122.0548	16.78	5.10
16	18 Apr 1990	13:53:51.30	36.9320	121.6577	5.68	5.75
17	18 Apr 1990	15:46:03.44	36.9583	121.6848	6.67	5.36
18	16 Jan 1993	06:29:35.00	37.0182	121.4630	7.90	5.10
19	11 Aug 1993	22:33:04.00	37.3120	121.6793	9.35	5.00
20	23 Apr 1995	08:41:36.63	36.6025	121.2015	7.84	5.14
21	12 Aug 1998	14:10:25.12	36.7533	121.4622	9.13	5.35
22	3 Sept 2000	08:36:30.09	38.3788	122.4127	10.12	5.17

Earthquakes in California

Wherever there are earthquakes there must be faults

Using seismic recording from around the world and global travel time curves, Earthquakes are located globally.



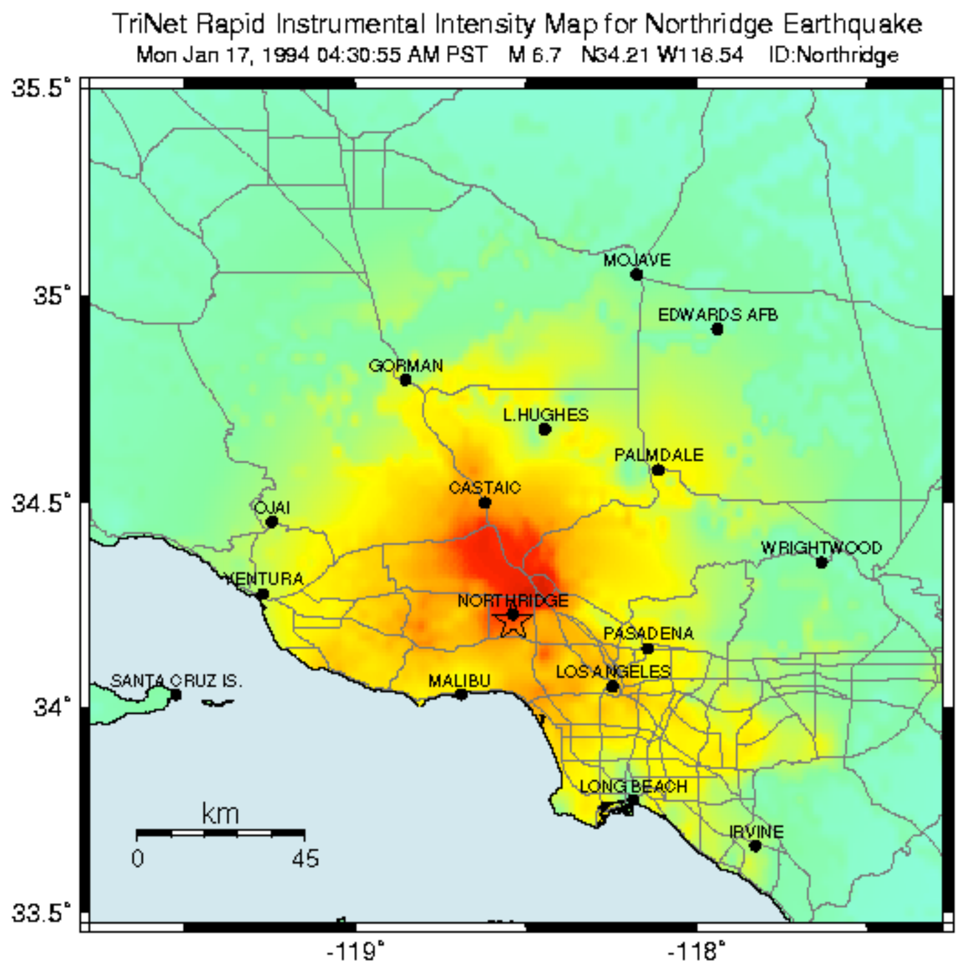

EARTHQUAKE DISTRIBUTION
(FOCAL DEPTHS 100 TO 700 km)

SEISMIC WAVES ATTENUATE - DECREASE IN AMPLITUDE - AS THEY PROPAGATE

Important for earth physics, understanding earthquake size, and seismic hazard

The wavefront area is expanding (fixed energy spreads over bigger area).

And, some energy is lost by non-elastic proceses.



PROCESSED: Tue Mar 6, 2001 11:35:21 AM PST,

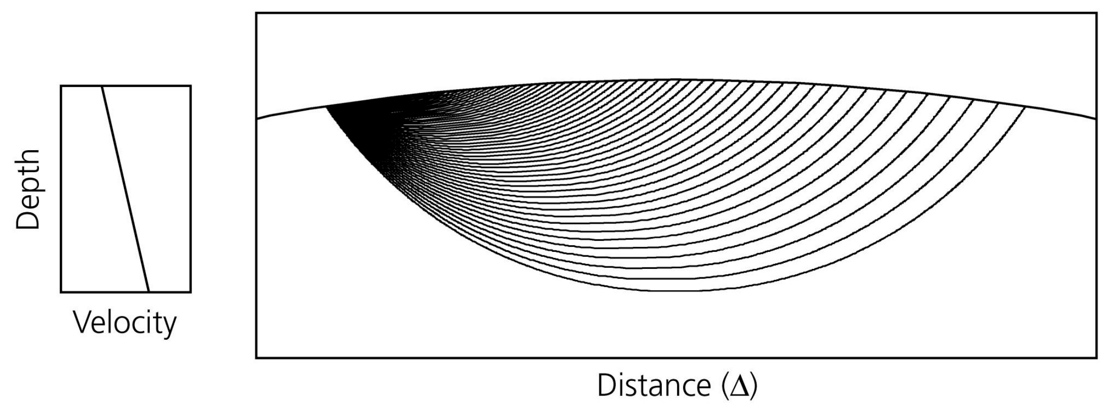
PERCEIVED SHAKING	Not felt	Weak	Light	Moderate	Strong	Very strong	Severe	Violent	Extreme
POTENTIAL DAMAGE	none	none	none	Very light	Light	Moderate	Moderate/Heavy	Heavy	Very Heavy
PEAK ACC.(%g)	<.17	.17-1.4	1.4-3.9	3.9-9.2	9.2-18	18-34	34-65	65-124	>124
PEAK VEL.(cm/s)	<0.1	0.1-1.1	1.1-3.4	3.4-8.1	8.1-16	16-31	31-60	60-116	>116
INSTRUMENTAL INTENSITY	I	II-III	IV	V	VI	VII	VIII	IX	X+

GEOMETRIC SPREADING: P and S BODY WAVES

For body waves, consider a spherical wavefront moving away from a deep earthquake. Energy is conserved on the expanding spherical wavefront whose area is $4 \pi r^2$, where r is the radius of the wavefront.

Thus the energy per unit wave front decays as $1 / r^2$, and the amplitude decreases as $1 / r$

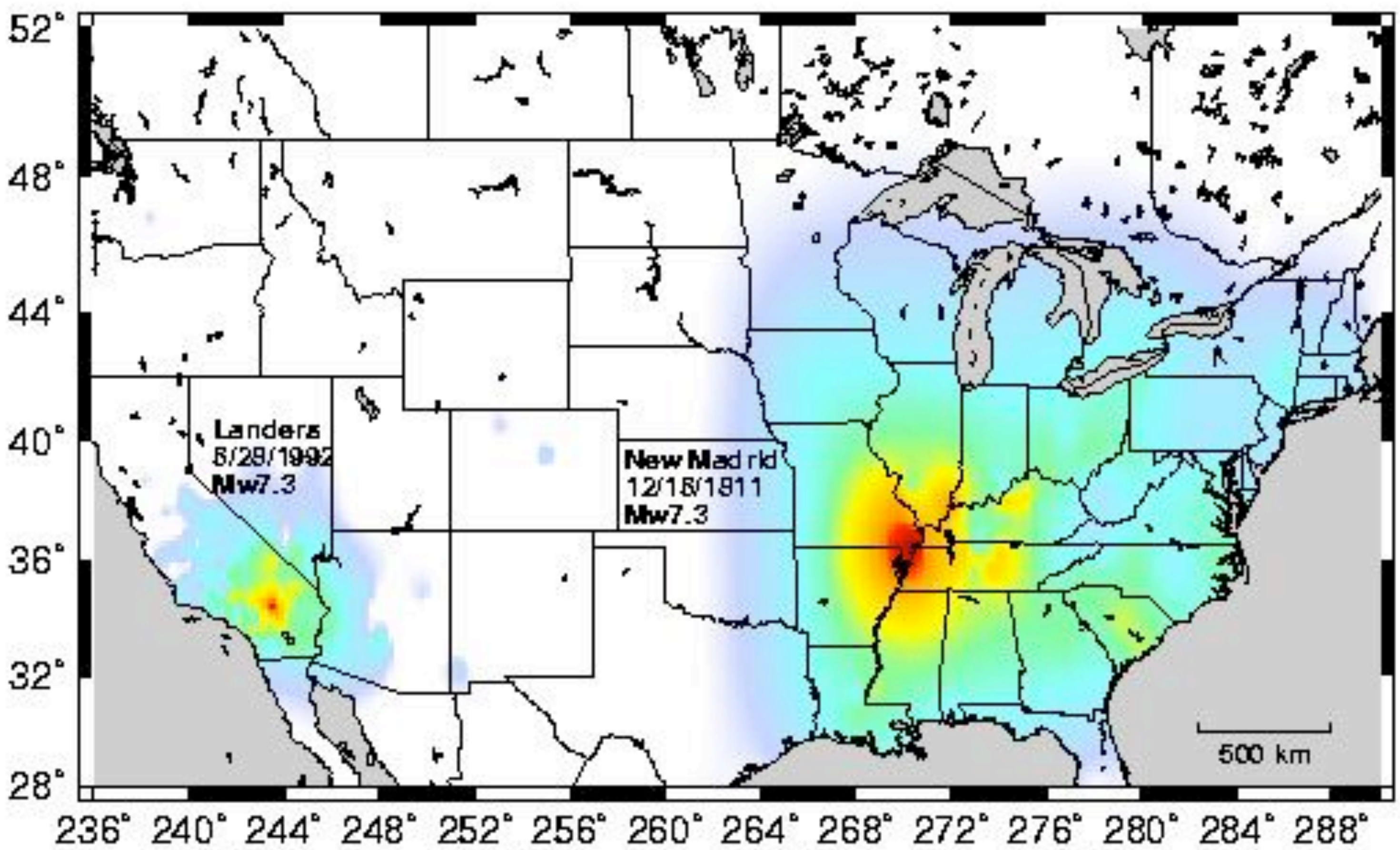
Figure 3.4-5: Ray path effects for increasing velocity.



In reality, because body waves travel through an inhomogeneous earth, their amplitude also depends on the focusing and defocusing of rays by the velocity structure.

Anelastic processes vary in the Lithosphere.

Seismic wave amplitudes decay more rapidly with distance in the western US.



PERCEIVED SHAKING	Not felt	Weak	Light	Moderate	Strong	Very strong	Severe	Violent	Extreme
POTENTIAL DAMAGE	none	none	none	Very light	Light	Moderate	Moderate/Heavy	Heavy	Very Heavy
PEAK ACC.(%g)	<.17	.17-1.4	1.4-3.9	3.9-9.2	9.2-18	18-34	34-65	65-124	>124
PEAK VEL.(cm/s)	<0.1	0.1-1.1	1.1-3.4	3.4-8.1	8.1-16	16-31	31-60	60-116	>116
INSTRUMENTAL INTENSITY	I	II-III	IV	V	VI	VII	VIII	IX	X+

http://pasadena.wr.usgs.gov/office/hough/east-vs-west.jpg

Richter magnitude scale ~ 1933

Richter magnitude = ML = log(amplitude/wave period) + distance correction

First of the "seismic magnitudes",

Strictly speaking, specific to a particular instrument type (Wood Anderson), in a particular region (southern California).

Can compare relative 'strength' of warthquakes based on how large the seismic waves are, corrected for propagation distance.

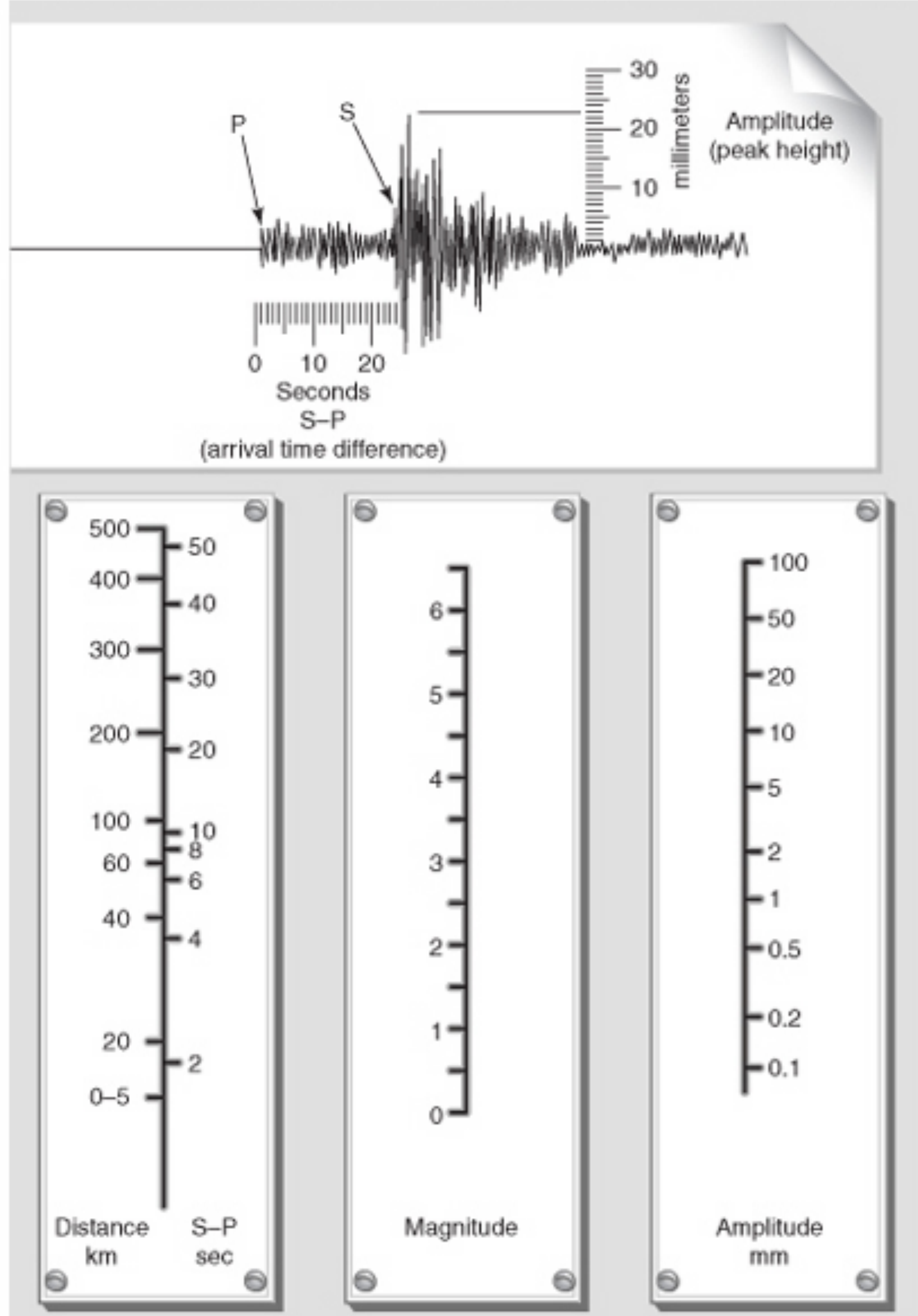
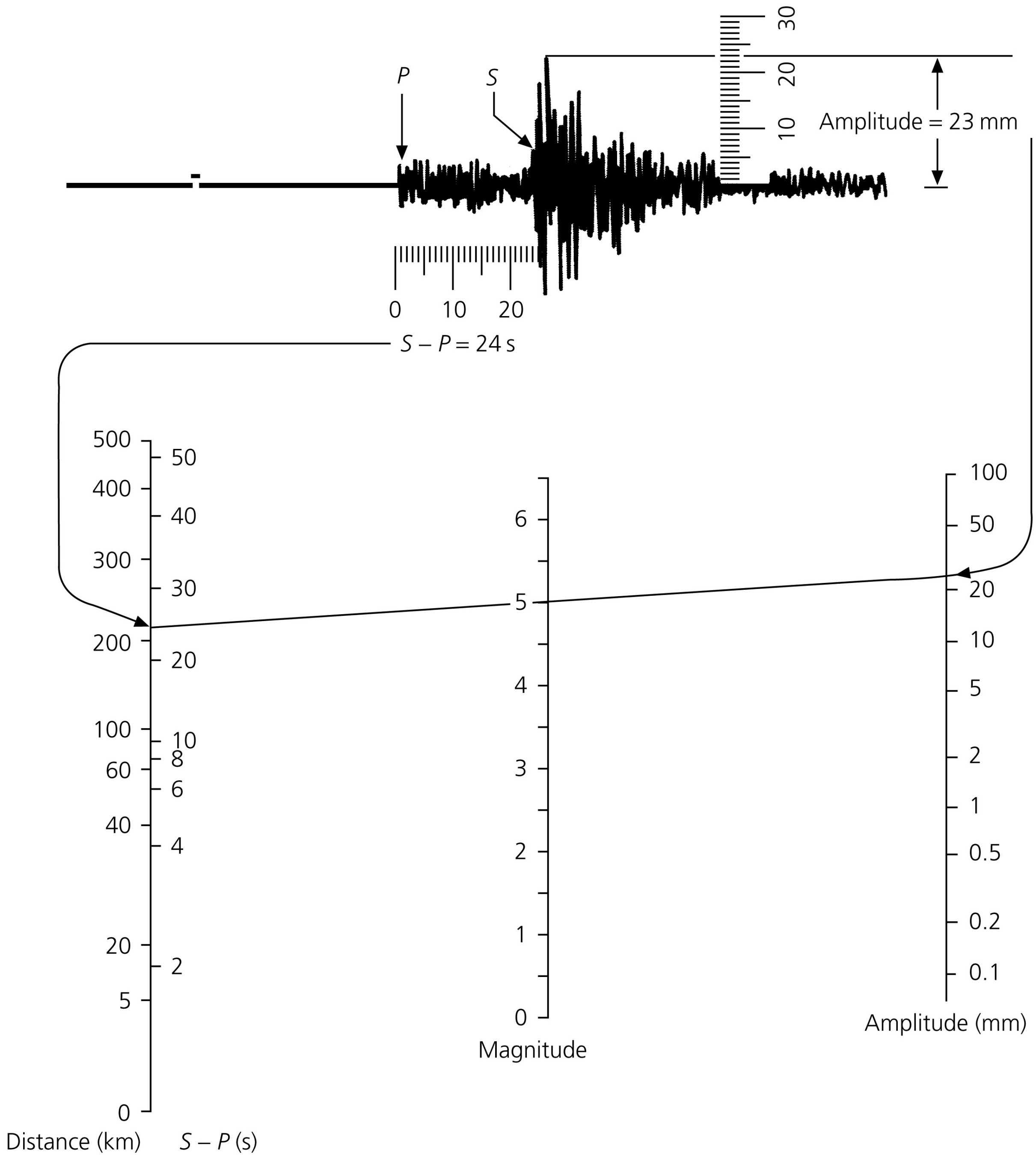


Figure 4.6-1: Example of the determination of the Richter scale.



- Supplement:

The seismic instrument produces a record that is modified from the actual ground motion. We have to determine the "instrument response" so that we know how to take the recorded seismogram and to relate it to actual ground shaking. Example calculation of the instrument response is shown in the next few slides. Instrument responses are calculated by theory and by direct shaking with a known input.

To solve it, we assume that

$$u(t) = e^{-i\omega t} \quad \text{and} \quad \xi(t) = X(\omega)e^{-i\omega t} \tag{4}$$

and substitute (4) into (3) to yield

$$X(\omega)(-\omega^2 - 2\varepsilon i\omega + \omega_0^2)e^{-i\omega t} = \omega^2 e^{-i\omega t} \tag{5}$$

or

$$X(\omega) = -\,\omega^2 \,/\, (\omega^2 - \omega_0^2 + 2\varepsilon i\omega), \tag{6}$$

which is the instrument response produced by a ground motion $e^{i\omega t}$.

$X(\omega)$ is complex and can be written in terms of the amplitude and phase responses

$$X(\omega) = |X(\omega)|e^{i\phi(\omega)}, \tag{7}$$

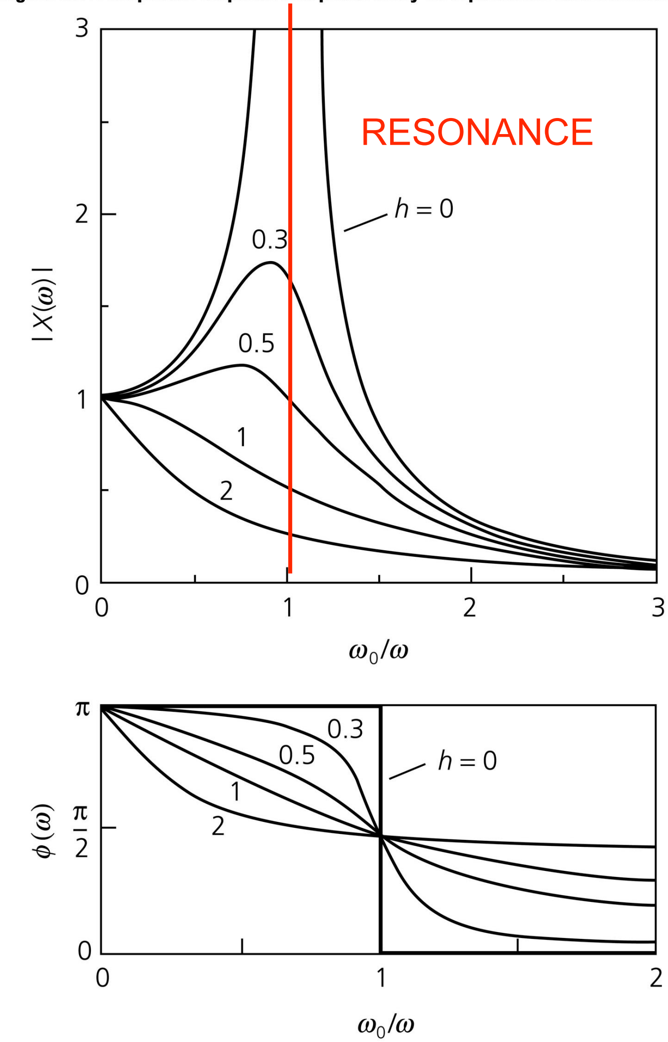
where

$$|X(\omega)| = \omega^2 \,/\, \left[(\omega^2 - \omega_0^2)^2 + 4\varepsilon^2\omega^2\right]^{1/2} \tag{8}$$

$$\phi(\omega) = -\tan^{-1}\frac{2\varepsilon\omega}{\omega^2 - \omega_0^2} + \pi. \tag{9}$$

As shown in Figure 2, these functions have several interesting features. First, as the angular frequency of the ground motion, ω, approaches the natural frequency of the pendulum, ω_0, the amplitude response is large. This effect, called *resonance,* is like "pumping" a playground swing at its natural period. Thus the seismometer responds best to ground motion near its natural period.

Figure 6.6-2: Amplitude response and phase delay for a pendulum seismometer.



ω_0 = (k/m)$^{1/2}$

Mass m

Spring constant k

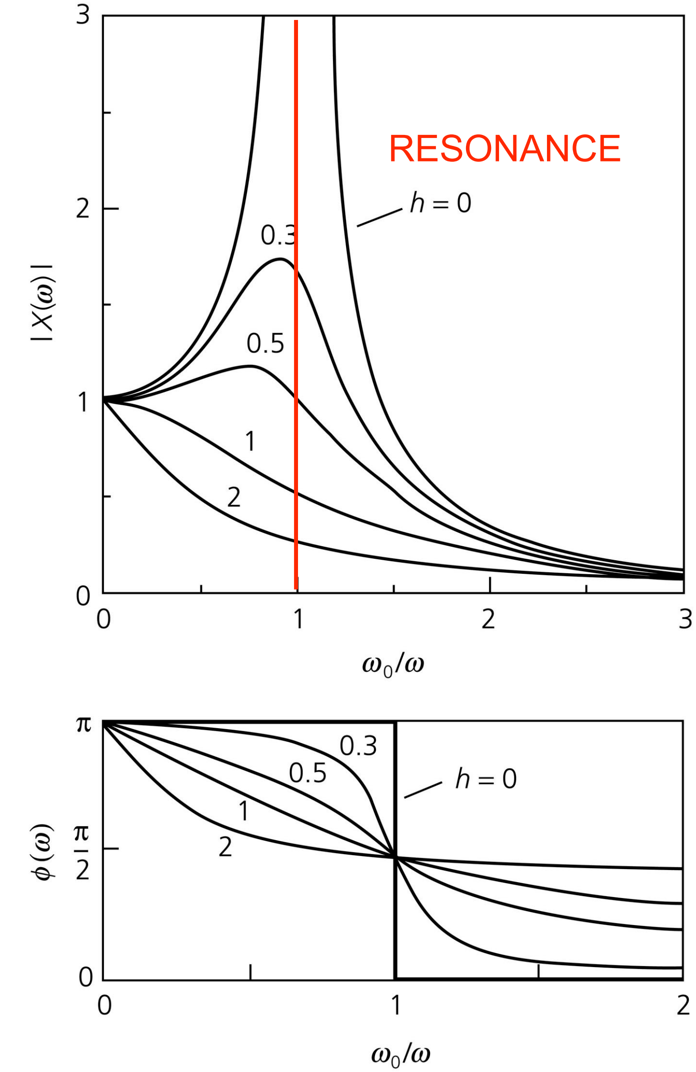
Damping h

For frequencies much greater than the natural frequency, $\omega \gg \omega_0$, $|X(\omega)| \to 1$, and $\phi(\omega) \to \pi$, so the seismometer records the ground motion but with the sign reversed. To see why this occurs consider (3): for $\omega \gg \omega_0$ the $\ddot{\xi}$ term is the largest term on the left side, so $\ddot{\xi}$ approximately equals $\ddot{u}$. Thus the seismometer responds to the ground *displacement*.

On the other hand, for frequencies much less than the natural frequency, $\omega \ll \omega_0$, $|X(\omega)| \to \omega^2/\omega_0^2$, and $\phi(\omega) \to 0$. Hence in this case the seismometer responds to *acceleration*, as can be seen from (3) because the $\omega_0^2\xi$ term is dominant and so ξ is proportional to $\ddot{u}$.

The shape of the instrument response depends on the damping factor $h = \varepsilon/\omega_0$. For $h = 0$, the system is undamped, and the amplitude response is peaked around the resonant frequency, $\omega = \omega_0$. The seismometer amplifi es ground motion with periods near its natural period. As damping is increased, the curve is smeared out. Thus the natural period and damping are used to design a seismometer to record ground motion in a particular period range.

Figure 6.6-2: Amplitude response and phase delay for a pendulum seismometer.



$\omega_0 = (k/m)^{1/2}$

Mass m

Spring constant k

Damping h

DIGITAL SEISMOMETER SYSTEM

Figure 6.6-12: Diagram showing the analog-to-digital (ADC) process.

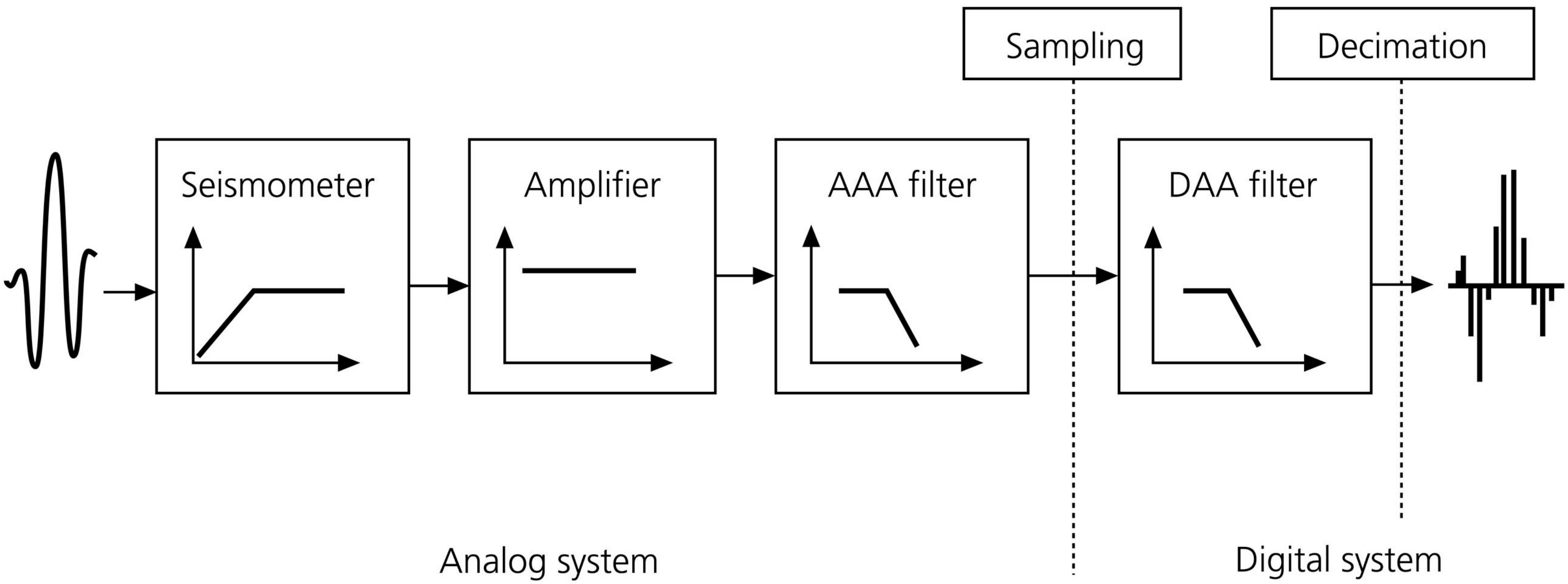


Figure 6.6-8: Instrument responses for several types of seismometers.

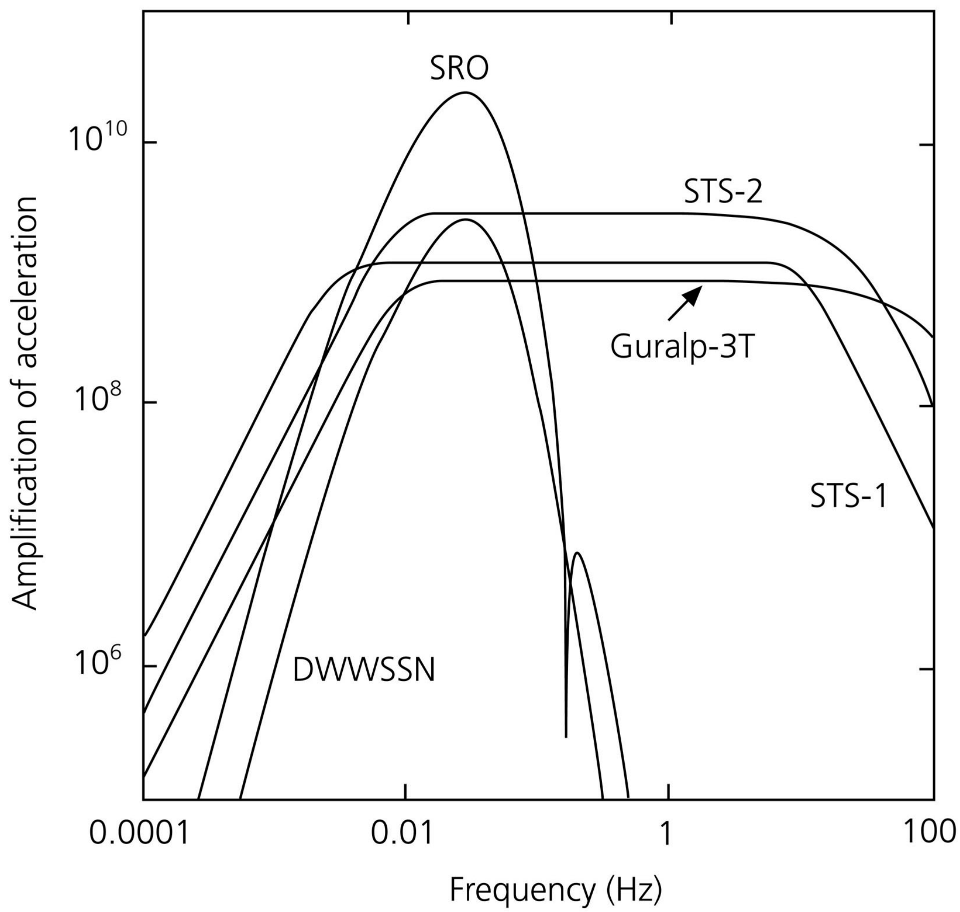


Figure 6.6-11: Use of filtering to enhance different frequency bands of a single seismogram.