PHS 314 SOLID EARTH PHYSICS (3 CU)

READING TEXT Fundamentals of Geophysics – William Lowrie

SEISMOLOGY

SEISMIC WAVES: Introduction

The propagation of a seismic disturbance through a heterogeneous medium is extremely complex. In order to derive equations that describe the propagation adequately, it is necessary to make simplifying assumptions. The heterogeneity of the medium is often modelled by dividing it into parallel layers, in each of which homogeneous conditions are assumed. By suitable choice of the thickness, density and elastic properties of each layer, the real conditions can be approximated. The most important assumption about the propagation of a seismic disturbance is that it travels by elastic displacements in the medium. This condition certainly does not apply close to the seismic source. In or near an earthquake focus or the shot point of a controlled explosion the medium is destroyed. Particles of the medium are displaced permanently from their neighbors; the deformation is anelastic. However, when a seismic disturbance has travelled some distance away from its source, its amplitude decreases and the medium deforms elastically to permit its passage. The particles of the medium carry out simple harmonic motions, and the seismic energy is transmitted as a complex set of wave motions.

When seismic energy is released suddenly at a point P near the surface of a homogeneous medium part of the energy propagates through the body of the medium as seismic *body waves.* The remaining part of the seismic energy spreads out over the surface as a seismic *surface wave,* analogous to the ripples on the surface of a pool of water into which a stone has been thrown.

Seismic body waves

When a body wave reaches a distance *r* from its source in a homogeneous medium, the *wavefront* (defined as the surface in which all particles vibrate with the same phase) has a spherical shape, and the wave is called a *spherical wave.* As the distance from the source increases, the curvature of the spherical wavefront decreases. At great distances from the source the wavefront is so flat that it can be considered to be a plane and the seismic wave is called a *plane wave.* The direction perpendicular to the wavefront is called the seismic *ray path.* The description of the harmonic motion in plane waves is simpler than for spherical waves, because for plane waves we can use orthogonal Cartesian coordinates. Even for plane waves the mathematical description of the three-dimensional displacements of the medium is fairly complex. However, we can learn quite a lot about body-wave propagation from a simpler, less rigorous description.

Compressional waves

Let Cartesian reference axes be defined such that the x-axis is parallel to the direction of propagation of the plane wave; the y- and z-axes then lie in the plane of the wavefront (Fig. 3.10). A generalized vibration of the medium can be reduced to components parallel to each of the reference axes. In the x-direction the particle motion is back and forward parallel to the direction of propagation. This results in the medium being alternately stretched and condensed in this direction. This harmonic motion produces a body wave that is transmitted as a sequence of rarefactions and condensations parallel to the x-axis.

Consider the disturbance of the medium shown. The area of the wavefront normal to the x-direction is A_x, and the wave propagation is treated as onedimensional. At an arbitrary position x, the passage of the wave produces a displacement u and a force F in the x-direction. At the position x+ dx the displacement is $u+du$ and the force is F_x+dF . Here dx is the infinitesimal length of a small volume element which has mass p dx A_x. The net force acting on this element in the x-direction is given by

$(F+dF)-F=dF=$ ax dx

The force F_X is caused by the stress element ox_x acting on the area A_x, and is equal to $o A_x$. This allows us to write the one-dimensional equation of motion

Kepler's laws of planetary motion

Kepler took many years to fit the observations of Tycho Brahe into three laws of planetary motion. The first and second laws were published in 1609 and the third law appeared in 1619. The laws may be formulated as follows:

1. the orbit of each planet is an ellipse with the Sun at one focus;
2. the orbital radius of a planet sweeps out equal areas in equal intervals of time;
3. the ratio of the square of a planet's period (T^2) to the cube of the semi-major axis of its orbit *(a^3)* is a constant for all planets.

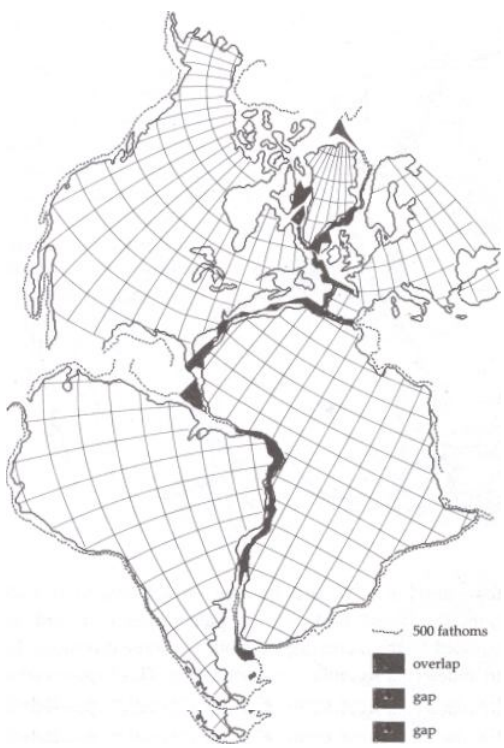
Bode's law

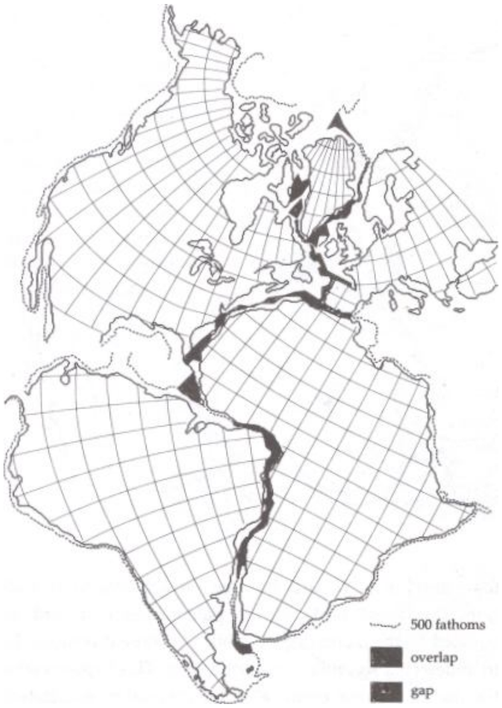
In 1772 the German astronomer Johann Bode devised an empirical formula to express the approximate distances of the planets from the Sun. A series of numbers is created in the following way: the first number is zero, the second is 0.3, and the rest are obtained by doubling the previous number. This gives the sequence 0, 0.3, 0.6, 1.2, 2.4, 4.8, 9.6, 19.2, 38.4, 76.8, etc. Each number is then augmented by 0.4 to give the sequence: 0.4, 0.7, 1.0, 1.6, 2.8, 5.2, 10.0, 19.6, 38.8, 77.2, etc. This series can be expressed mathematically as follows:

d,=0.4 for n= *d=0.4+0.3 - 0 . 4 + 0 . 3 x 2 " - 2* for $n=2^n$

This expression gives the distance $d_\circ$ in astronomical units (AU) of the nth planet from the Sun. It is usually known as Bode's law, but, as the same relationship had been suggested earlier by J. D. Titius of Wittenberg, it is sometimes called Titius-Bode's law.

Fits of the Continents: Pangaea and Gondwannaland

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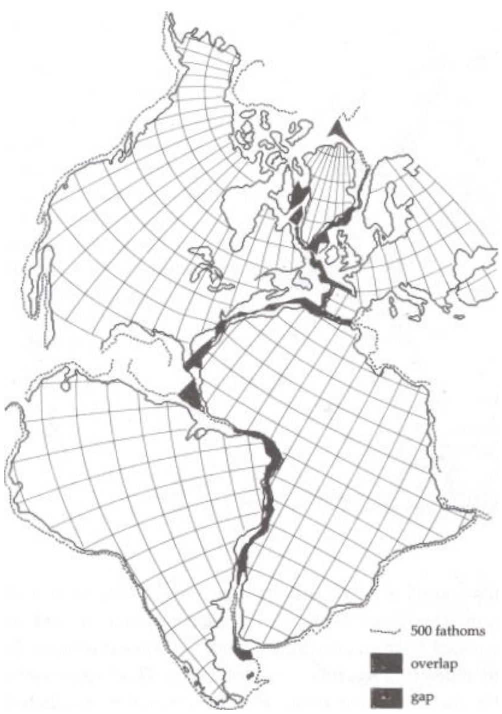
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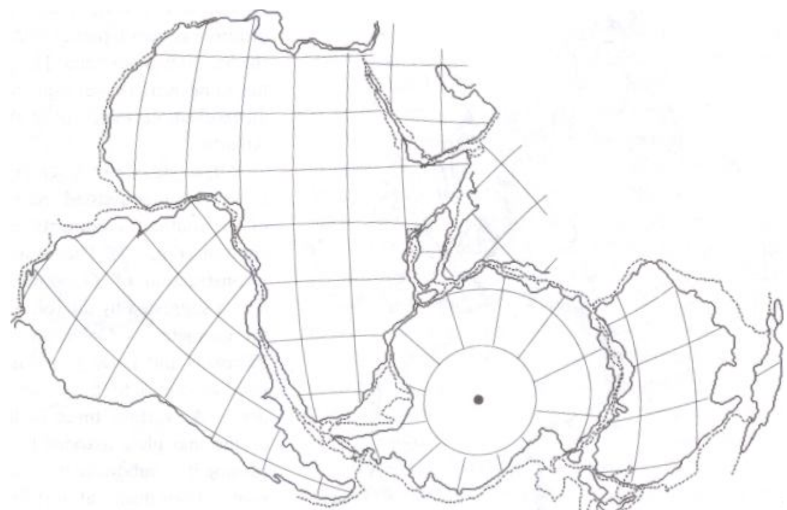


Structure of the Earth

Simplified layered structure of the Earth's interior showing the depths of the most important seismic discontinuities /.

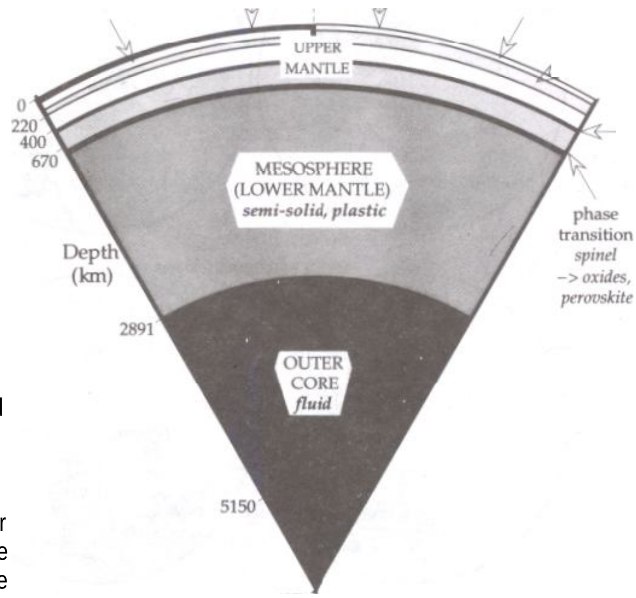
CONTINENT OCEAN

LITHOSPHERE Crust Crust LITHOSPHERE
rigid *38-40* km thick *6-8* km thick *rigid*
100-150 km thick *70-100* km thick

ASTHENOSPHERE

/partially molten

phase transition olivine -> spine!



The Earth has a radially layered structure, like that of an onion The boundaries between the layers are marked by abrupt changes in seismic velocity or velocity gradient. Each layer is characterized by a specific set of physical properties determined by the composition, pressure and temperature in the layer. The four main layers are i) the crust, ii) mantle and iii(the outer and iv) inner cores.

Lithospheric plates

The radially layered model of the Earth's interior assumes spherical symmetry. This is not valid for the crust and upper mantle. These outer layers of the Earth show important lateral variations. The crust and uppermost mantle down to a depth of about 70-100 km under deep ocean basins and 100-150 km under continents is rigid, forming a hard outer shell called the *lithosphere.* Beneath the lithosphere lies the *asthenosphere,* a layer in which seismic velocities often decrease, suggesting lower rigidity. It is about 150 km thick, although its upper and lower boundaries are not sharply defined. This weaker layer is thought to be partially molten; it may be able to flow over long periods of time like a viscous motions of the overlying lithospheric plates.

The brittle condition of the lithosphere causes it to fracture when strongly stressed. The rupture produces an earthquake, which is the violent release of elastic energy due to sudden displacement on a fault plane. Earthquakes are not distributed evenly over the surface of the globe, but occur predominantly in well-defined narrow seismic zones that are often associated with volcanic activity. These are: (a) the circum-Pacific `ring of fire'; (b) a sinuous belt running from the Azores through North Africa and the Alpine-Dinaride-Himalayan mountain chain as far as S.E. Asia; and (c) the world-circling system of oceanic ridges and rises. The seismic zones subdivide the lithosphere laterally into *tectonic plates*. A plate may be as broad as 10,000 km (e.g., the Pacific plate) or as small as a few thousand km (e.g., the Philippines plate). There are twelve major plates (Antarctica, Africa, Eurasia, India, Australia, Arabia, Philippines, North America, South America, Pacific, Nazca, and Cocos) and several minor plates (e.g., Scotia, Caribbean, Juan de Fuca). The positions of the boundaries between the North American and South American plates and between the North American and Eurasian plates are uncertain. The boundary between the Indian and Australian plates is not sharply defined, but may be a broad region of diffuse deformation.

Types of plate margin

An important factor in the evolution of modern plate tectonic theory was the development of oceanography in the years following World War II, when technology designed for warfare was turned to peaceful purposes. The bathymetry of the oceans was charted extensively by echo-sounding and within a few years several striking features became evident. Deep trenches, more than twice the depth of the ocean basins, were discovered close to island arcs and some continental margins; the Marianas Trench is more than 11 km deep. A prominent submarine mountain chain - called an oceanic ridge -

was found in each ocean. The oceanic ridges rise to as much as 3000 m above the adjacent basins and form a continuous system, more than 60,000 km in length, that girdles the globe. Unlike continental mountain belts, which are usually less than several hundred kilometers across, the oceanic ridges are 2000-4000 km in width. The ridge system is offset at intervals by long horizontal faults forming fracture zones. These three features - trenches, ridges and fracture zones - originate from different plate tectonic processes.

The lithospheric plates are very thin in comparison to their breadth. Most earthquakes occur at plate margins, and are associated with interactions between plates. Apart from rare intraplate earthquakes, which can be as large and disastrous as the earthquakes at plate boundaries, the plate interiors are aseismic. This suggests that the plates behave rigidly. Analysis of earthquakes allows the direction of displacement to be determined and permits interpretation of the relative motions between plates.

There are three types of plate margin, distinguished by different tectonic processes. These are:

i) Constructive plate margin; ii) Destructive plate margin; and iii) Conservative plate margin.

The world-wide pattern of earthquakes shows that the plates are presently moving apart at oceanic ridges. Magnetic evidence, discussed below, confirms that the separation has been going on for millions of years. New lithosphere is being formed at these spreading centers, so the ridges can be regarded as con structive plate margins. The seismic zones related to deepsea trenches, island arcs and mountain belts mark places where lithospheric plates are converging. One plate is forced under another there in a so-called subduction zone. Because it is thin in relation to its breadth, the lower plate bends sharply before descending to depths of several hundred kilometers, where it is absorbed. The subduction zone marks a destructive plate margin.

Constructive and destructive plate margins may consist of many segments linked by horizontal faults. A crucial step in the development of plate tectonic theory was made in 1965 by a Canadian geologist, J. Tuzo Wilson, who recognized that these faults are not conventional transcurrent faults. They belong to a new class of faults, which Wilson called transform faults. The relative motion on a transform fault is opposite to what might be inferred from the offsets of bordering ridge segments. At the point where a transform fault meets an oceanic ridge it transforms the spreading on the ridge to horizontal shear on the fault. Likewise, where such a fault meets a destructive plate margin it transforms subduction to horizontal shear.

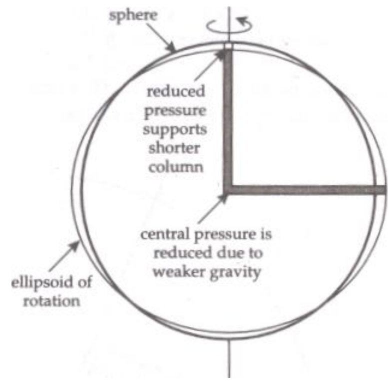
The transform faults form a conservative plate margin, where lithosphere is neither created nor destroyed; the boundary separates plates that move past each other horizontally. Earthquake activity on an oceanic ridge system was confined almost entirely to the transform fault between ridge crests, where the neighboring plates rub past each other; the mechanisms of earthquakes on the transform faults agreed with the predicted sense of strike-slip motion.

Transform faults play a key role in determining plate motions. Spreading and subduction are often assumed to be perpendicular to the strike of a ridge or trench. This is not necessarily the case. Oblique motion with a component along strike is possible at each of these margins. However, because lithosphere is neither created nor destroyed at a conservative margin, the relative motion between adjacent plates must be parallel to the strike of a shared transform fault.

Shape of the Earth

The Earth's radius was earljer calculated to be 6372 km, remarkably close to the modern value of 6371

In 1672 French astronomer, Jean Richer, was sent by Louis XIV to make astronomical observations on the equatorial island of Cayenne. He found that an accurate pendulum clock, which had been adjusted in Paris precisely to beat seconds, was losing about two and a half minutes per day, i.e. its period was now too long. The error was much too large to be explained by inaccuracy of the precise instrument. The observation aroused much interest and speculation, but was only explained some 15 years later by Sir Isaac Newton in terms of his laws of universal gravitation and motion.



Newton's argument that the shape of the rotating Earth should be flattened at the poles and bulge at the equator was based on hydrostatic equilibrium between polar and equatorial pressure columns (after Strahler, 1963).

Newton argued that the shape of the rotating Earth should be that of an *oblate ellipsoid;* compared to a sphere, it should be somewhat flattened at the poles and should bulge outward around the equator. This inference was made on logical grounds. Assume that the Earth does not rotate and that holes could be drilled to its center along the rotation axis and along an equatorial radius. If these holes are filled with water, the hydrostatic pressure at the center of the Earth sustains equal water columns along each radius. However, the rotation of the Earth causes a centrifugal force at the equator but has no effect on the axis of rotation. At the equator the outward centrifugal force of the rotation opposes the inward gravitational attraction and pulls the water column upward. At the same time it reduces the hydrostatic pressure produced by the water column at the Earth's center. The reduced central pressure is unable to support the height of the water column along the polar radius, which subsides. If the Earth were a hydrostatic sphere, the form of the rotating Earth should be an oblate ellipsoid of revolution. Newton assumed the Earth's density to be constant and calculated that the flattening should be about 1:230 (roughly 0.5%). This is somewhat larger than the actual flattening of the Earth, which is about 1:298 (roughly 0.3%).

The increase in period of Richer's pendulum could now be explained. Cayenne was close to the equator, where the larger radius placed the observer further from the center of gravitational attraction, and the increased distance from the rotational axis resulted in a stronger opposing centrifugal force. These two effects resulted in a lower value of gravity in Cayenne than in Paris, where the clock had been calibrated.

There was no direct proof of Newton's interpretation. A corollary of his interpretation was that the degree of meridian arc should subtend a longer distance in polar regions than near the equator Early in the 18th century French geodesists extended the standard meridian from border to border of the country and found a puzzling result. In contrast to the prediction of Newton, the degree of meridian arc *decreased* northward. The French interpretation was that the Earth's shape was *a prolate* ellipsoid, elongated at the poles and narrowed at the equator, like the shape of a rugby football. A major scientific controversy arose between the `flatteners' and the 'elongators'.

To determine whether the Earth's shape is oblate or prolate, the Academie Royale des Sciences sponsored two scientific expeditions. In 1736-1737 a team of scientists measured the length of a degree of meridian arc in Lapland, near the Arctic Circle. They found a length appreciably longer than the meridian degree measured by Picard near Paris. From 1735 to 1743 a second party of scientists measured the length of more than 3 degrees of meridian arc in Peru, near the equator. Their results showed that the equatorial degree of latitude was shorter than the meridian degree in Paris. Both parties confirmed convincingly the prediction of Newton that the Earth's shape is that of an oblate ellipsoid.

The ellipsoidal shape of the Earth resulting from its rotation has important consequences, not only for the variation with latitude of gravity on the Earth's surface, but also for the Earth's rate of rotation and the orientation of its rotational axis. These are modified by torques that arise from the gravitational attractions of the Sun, Moon and planets on the ellipsoidal shape.

THE TIDES

The gravitational forces of Sun and Moon deform the Earth's shape, causing tides in the oceans, atmosphere and solid body of the Earth. The most visible tidal effects are the displacements of the ocean surface, which is a hydrostatic equipotential surface. The Earth does not react rigidly to the tidal forces. The solid body of the Earth deforms in a like manner to the free surface, giving rise to so-called *bodily Earth-tides.* These can be observed with specially designed instruments, which operate on a similar principle to the long-period seismometer. The height of the marine equilibrium tide amounts to only half a meter or so over the free ocean. In coastal areas the tidal height is significantly increased by the shallowing of the continental shelf and the confining shapes of bays and harbors. Accordingly, the height and variation of the tide at any place is influenced strongly by complex local factors. Subsequent chapters deal with the tidal deformations of the Earth's hydrostatic figure.

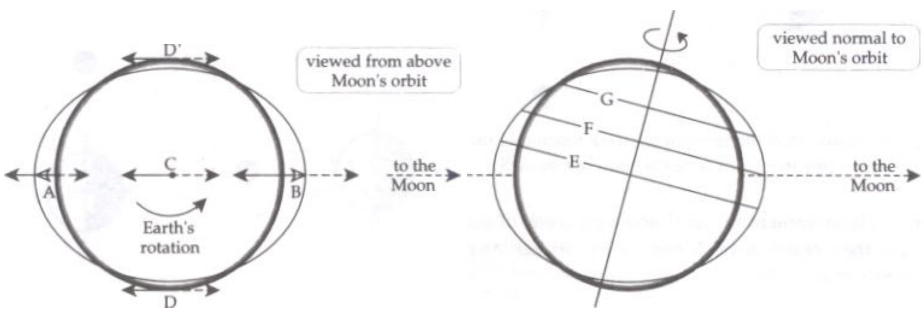
Lunar tidal periodicity

The figure illustrates the principle of the `revolution without rotation 'of the Earth-Moon pair about their common center of mass at S.

The Earth and Moon are coupled together by gravitational attraction. Their common motion is like that of a pair of ballroom dancers. Each partner moves around the center of mass of the pair. For the Earth-Moon pair the location of the center of mass is easily found. Let *E* be the mass of the Earth, and *m* that of the Moon; let the separation of the centers of the Earth and Moon be r_L and let the distance of their common center of mass be *d* from the center of the Earth. The moment of the Earth about the center of mass is *Ed* and the moment of the Moon is $m(r_L\text{-}d)$. Setting these moments equal we get

The mass of the Moon is 0.0123 that of the Earth and the distance between the centers is 382,000 km. These figures give *d=4600* km, i.e., the center of revolution of the Earth-Moon pair lies within the Earth.

To understand the common revolution of the Earth-Moon pair we have to exclude the rotation of the Earth about its axis. The revolution without rotation' is illustrated in the figure. The Earth-Moon pair revolves about S, the center of mass. Let the starting positions be as shown in fig. a. Approximately one week later the Moon has advanced in its path by one-quarter of a revolution and the center of the Earth has moved so as to keep the center of mass fixed (fig. b). The relationship is maintained in the following weeks (figs. c, d) so that during one month the center of the Earth describes a circle about S. Now consider the motion of point number 2 on the left hand side of the Earth in the figure. If the Earth revolves as a rigid body and the rotation about its own axis is omitted, after one week point 2 will have moved to a new position but will still be the furthest point on the left. Subsequently, during one month point 2 will describe a small circle with the same radius as the circle described by the Earth's center. Similarly points 1,



What is revolution without rotation?

The `revolution without rotation' causes each point in the body of the Earth to describe a circular path with identical radius. The centrifugal acceleration of this motion has therefore the same magnitude at all points in the Earth, and, it is directed away from the Moon parallel to the Earth-Moon line of centers.

What happens at C?

At C, the center of the Earth, this centrifugal acceleration exactly balances the gravitational attraction of the Moon.

What happens at B?

At B, on the side of the Earth nearest to the Moon, the gravitational acceleration of the Moon is larger than at the center of the Earth and exceeds the centrifugal acceleration a_L. There is a residual acceleration *toward the* Moon, which raises a tide on this side of the Earth.

What happens at A?

At A, on the far side of the Earth, the gravitational acceleration of the Moon is less than the centrifugal acceleration a_L. The residual acceleration is *away from* the Moon, and raises a tide on the far side of the Earth.

What happens at D and D"?

At points D and D' the direction of the gravitational acceleration due to the Moon is not exactly parallel to the line of centers of the Earth-Moon pair. The residual tidal acceleration is almost along the direction toward the center of the Earth. Its effect is to lower the free surface in this direction.

Can you use this to predict the shape of the Earth?

The tidal deformation of the Earth produced by the Moon has a prolate ellipsoidal shape, like a rugby football, along the Earth-Moon line of centers. The daily tides are caused by superposing the Earth's rotation on this deformation. In the course of one day a point rotates past the points A, D, B and D' and an observer experiences two full tidal cycles, called the *semi-diurnal* tides. The extreme tides are not equal at every latitude, because of the varying angle between the Earth's rotational axis and the Moon's orbit. At the equator E the semi-diurnal tides are equal; at an intermediate latitude F one tide is higher than the other; and at latitude G and higher there is only one (diurnal) tide per day. The difference in height between two successive high or low tides is called the *diurnal inequality.*

Has the Sun any influence on the tides?

The Sun also has an influence on the tides. The theory of the solar tides can be followed in identical manner to the lunar tides by again applying the principle of `revolution without rotation'. The Sun's mass is 333,340 times greater than that of the Earth, so the common center of mass is close to the center of the Sun at a radial distance of about 450 km from its center. The period of the revolution is one year. As for the lunar tide, the imbalance between gravitational acceleration of the Sun and centrifugal acceleration due to the common revolution leads to a prolate ellipsoidal tidal deformation. The solar effect is smaller than that of the Moon. Although the mass of the Sun is vastly greater than that of the Moon, its distance from the Earth is also much greater, and, because gravitational acceleration varies inversely with the square of distance, the maximum tidal effect of the Sun is only about 45% that of the Moon.

Spring and neap tides

The superposition of the lunar and solar tides causes a modulation of the tidal amplitude. The ecliptic plane is defined by the Earth's orbit around the Sun. The Moon's orbit around the Earth is not exactly in the ecliptic but is inclined at a very small angle of about 5.2° to it. For discussion of the combination of lunar and solar tides we can assume the orbits to be coplanar. The Moon and Sun each produce a prolate tidal deformation of the Earth, but the relative orientations of these ellipsoids vary during one month. At conjunction the (new) Moon is on the same side of the Earth as the Sun, and the ellipsoidal deformations augment each other. The same is the case half a month later at opposition, when the (full) Moon is on the opposite side of the Earth from the Sun. The unusually high tides at opposition and conjunction are called *spring tides.* In contrast, at the times of quadrature the waxing or waning half Moon causes a prolate ellipsoidal deformation out of phase with the solar deformation. The maximum lunar tide coincides with the minimum solar tide, and the effects partially cancel each other. The unusually low tides at quadrature are called *neap tides.* The superposition of the lunar and solar tides causes modulation of the tidal amplitude during a month.

Effect of the tides on gravity measurements

The tides have an effect on gravity measurements made on the Earth. The combined effects of Sun and Moon cause an acceleration at the Earth's surface of up to about 0.3 mgal, of which about two-thirds are due to the Moon and onethird to the Sun. The sensitive modern instruments used for gravity exploration can readily detect gravity differences of 0.01 mgal. It is necessary to compensate gravity measurements for the tidal effects, which vary with location, date and time of day. Fortunately, tidal theory is so well established that the gravity effect can be calculated and tabulated for any place and time before beginning a survey.

Bodily Earth-tides

A simple way to measure the height of the marine tide might be to fix a stake to the sea-bottom at a suitably sheltered location and to record continuously the measured water level (assuming that confusion introduced by wave motion can be eliminated or taken into account). The observed amplitude of the marine tide, defined by the displacement of the free water surface, is found to be about 70% of the theoretical value. The difference is explained by the elasticity of the Earth. The tidal deformation corresponds to a redistribution of mass, which modifies the gravitational potential of the Earth and augments the elevation of the free surface. This is partially counteracted by a bodily tide in the solid Earth, which deforms elastically in response to the attraction of the Sun and Moon. The free water surface is raised by the tidal attraction, but the sea-bottom in which the measuring rod is implanted is also raised. The measured tide is the difference between the marine tide and the bodily Earthtide.
In practice, the displacement of the equipotential surface is measured with a horizontal pendulum, which reacts to the tilt of the surface. The bodily Earth-tides also affect gravity measurements and can be observed with sensitive gravimeters. The effects of the bodily Earth-tides are incorporated into the predicted tidal corrections to gravity measurements.

Seismic Propagation

Introduction

The propagation of a seismic disturbance through a heterogeneous medium is extremely complex. In order to derive equations that describe the propagation adequately, it is necessary to make simplifying assumptions. The heterogeneity of the medium is often modelled by dividing it into parallel layers, in each of which homogeneous conditions are assumed. By suitable choice of the thickness, density and elastic properties of each layer, the real conditions can be approximated. The most important assumption about the propagation of a seismic disturbance is that it travels by elastic displacements in the medium. This condition certainly does not apply close to the seismic source. In or near an earthquake focus or the shot point of a controlled explosion the medium is destroyed. Particles of the medium are displaced permanently from their neighbors; the deformation is anelastic. However, when a seismic disturbance has travelled some distance away from its source, its amplitude decreases and the medium deforms elastically to permit its passage. The particles of the medium carry out simple harmonic motions, and the seismic energy is transmitted as a complex set of wave motions.

When seismic energy is released suddenly at a point P near the surface of a homogeneous medium, part of the energy propagates through the body of the medium as seismic *body waves.* The remaining part of the seismic energy spreads out over the surface as a seismic *surface wave,* analogous to the ripples on the surface of a pool of water into which a stone has been thrown.

Seismic body waves

When a body wave reaches a distance *r* from its source in a homogeneous medium, the *wavefront* (defined as the surface in which all particles vibrate with the same phase) has a spherical shape, and the wave is called a *spherical wave.* As the distance from the source increases, the curvature of the spherical wavefront decreases. At great distances from the source the wavefront is so flat that it can be considered to be a plane and the seismic wave is called a *plane wave.* The direction perpendicular to the wavefront is called the seismic *ray path.* The description of the harmonic motion in plane waves is simpler than for spherical waves, because for plane waves we can use orthogonal Cartesian coordinates. Even for plane waves the mathematical description of the three-dimensional displacements of the medium is fairly complex. However, we can learn quite a lot about body-wave propagation from a simpler, less rigorous description.

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Consider the disturbance of the medium shown. The area of the wavefront normal to the x-direction is A_x, and the wave propagation is treated as onedimensional. At an arbitrary position x, the passage of the wave produces a displacement u and a force F in the x-direction. At the position x+ dx the displacement is $u + du$ and the force is F_x+dF . Here dx is the infinitesimal length of a small volume element which has mass p dx A_x. The net force acting on this element in the x-direction is given by

$(F+dF)-F=dF=$ ax dx

The force F_X is caused by the stress element ox_x acting on the area A_x, and is equal to $o A_x$. This allows us to write the one-dimensional equation of motion

Longitudinal Waves

The longitudinal wave is the fastest of all seismic waves. When an earthquake occurs, this wave is the first to arrive at a recording station. As a result it is called the *primary wave,* or *P-wave.* Equation shows that P-waves can travel through solids, liquids and gases, all of which are compressible (K#0). Liquids and gases do not allow shear. Consequently, μ=0, and the compressional wave velocity in a fluid is given by

$$\alpha = \sqrt{\frac{K}{\rho}}$$

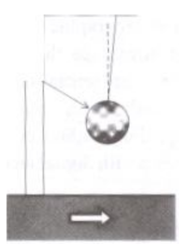
Transverse waves

The vibrations along they- and z-axes are parallel to the wavefront and transverse to the direction of propagation. If we wish, we can combine the y- and z-components into a single transverse motion. It is more convenient, however, to analyze the motions in the vertical and horizontal planes separately. Here we discuss the disturbance in the vertical plane defined by the x- and z-axes; an analogous description applies to the horizontal plane.

The transverse wave motion is akin to that seen when a rope is shaken. Vertical planes move up and down and adjacent elements of the medium experience shape distortions, changing repeatedly from a rectangle to a parallelogram and back. Adjacent elements of the medium suffer vertical shear.

THE SEISMOGRAPH, SEISMOMETER, AND SEISMOGRAM

The earliest known instrument for indicating the arrival of a seismic tremor from a distant source is reputed to have been invented by a Chinese astronomer called Chang Heng in 132 A.D. The device consisted of eight inverted dragons placed at equal intervals around the rim of a vase. Under each dragon sat an open-mouthed metal toad. Each dragon held a bronze ball in its mouth. When a slight tremor shook the device, an internal mechanism opened the mouth of one dragon, releasing its bronze ball, which fell into the open mouth of the metal toad beneath, thereby marking the direction of arrival of the tremor. The principle of this instrument was used in 18th century European devices that consisted of brimful bowls of water or mercury with grooved rims under which tiny collector bowls were placed



ground moves to left ground moves to right

The seismometer

Because of its inertia, a suspended heavy mass remains almost stationary when the ground and suspension move to the left or to the right. to collect the overflow occasioned by a seismic tremor. These instruments gave visible evidence of a seismic event but were unable to trace a permanent record of the seismic wave itself. They are classified as *seismoscopes.*

The science of seismology dates from the invention of the *seismograph* by the English scientist John Milne in 1892. Its name derives from its ability to convert an unfelt ground vibration into a visible record. The seismograph consists of a receiver and a recorder. The ground vibration is detected and amplified by a sensor, called the *seismometer* or, in exploration seismology, the *geophone.* In modern instruments the vibration is amplified and filtered electronically. The amplified ground motion is converted to a visible record, called the *seismogram*.

The seismometer makes use of the principle of inertia. If a heavy mass is only loosely coupled to the ground (for example, by suspending it from a wire like a pendulum as in the motion of the Earth caused by a seismic wave is only partly transferred to the mass. While the ground vibrates, the inertia of the heavy mass assures that it does not move as much, if at all. The seismometer amplifies and records the relative motion between the mass and the ground.

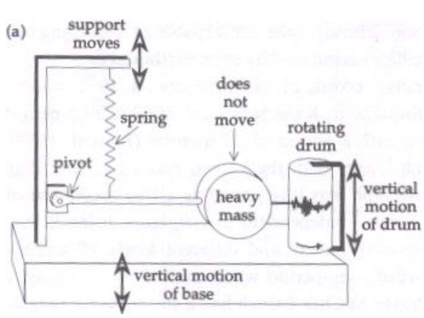
Principle of the seismometer

Seismometers are designed to react to motion of the Earth in a given direction. Mechanical instruments record the amplified displacement of the ground; electromagnetic instruments respond to the velocity of ground motion. Depending on the design, either type may respond to vertical or horizontal motion. Some modern electromagnetic instruments are constructed so as to record simultaneously three orthogonal components of motion. Most designs employ variations on the pendulum principle.

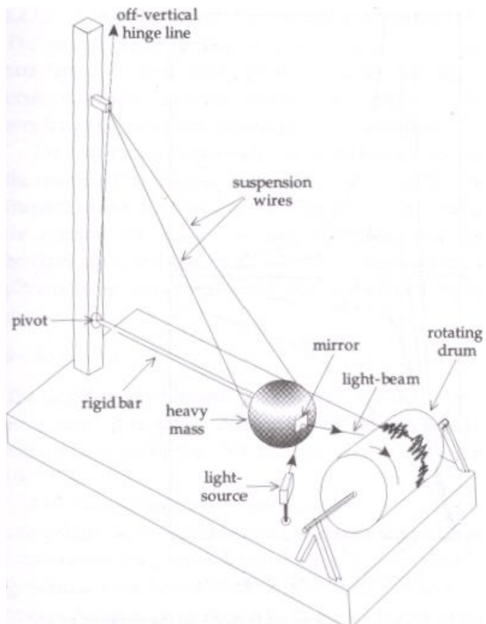
Vertical-motion seismometer

In the mechanical type of vertical-motion seismometer, a large mass is mounted on a horizontal bar hinged at a pivot so that it can move only in the vertical plane. A pen attached to the bar writes on a horizontal rotating drum that is fixed to the housing of the instrument. The bar is held in a horizontal position by a weak spring. This assures a loose coupling between the mass and the housing, which is connected rigidly to the ground. Vertical ground motion, as sensed during the passage of a seismic wave, is transmitted to the housing but not to the inertial mass and the pen, which remain stationary. The pen inscribes a trace of the vertical vibration of the housing on a paper fixed to the rotating drum. This trace is the vertical motion seismogram of the seismic wave.

The electromagnetic seismometer responds to the relative motion between a magnet and a coil of wire. One of these members is fixed to the housing of the instrument and thereby to the Earth. The other is suspended by a spring and forms the inertial member. Two basic designs are possible. In the moving-magnet type, the coil is fixed to the housing and the magnet is inertial. In the moving-coil type the roles are reversed. A coil of wire fixed to the inertial mass is suspended between the poles of a strong magnet, which in turn is fixed to the ground by the rigid housing. Any motion of the coil within the magnetic field induces a voltage in the coil proportional to the rate of change of magnetic flux. During a seismic arrival the vibration of the ground relative to the mass is converted to an electrical voltage by induction in the coil. The voltage is amplified and transmitted through an electrical circuit to the recorder with the exception that the axis of the moving member (coil or magnet) is horizontal.



Schematic diagrams illustrating the principle of operation of the vertical-motion seismometer: (a) mechanical pendulum type (after Strahler, 1963), (b) electromagnetic, moving coil type.

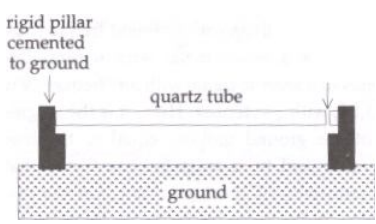


Schematic design of the pendulum type of horizontalmotion seismometer (after Strahler, 1963).

Strain seismometer

The pendulum seismometers described above are inertial devices, which depend on the resistance of a loosely coupled mass to a change in its momentum. At about the same time that he developed the inertial seismograph, Milne also conducted experiments with a primitive strain seismograph that measured the change in distance between two posts during the passage of a seismic wave. The gain of early strain seismographs was low. However, in *1935* H. Benioff invented a sensitive strain seismograph from which modern versions are descended.

The principle of the instrument is as shown. It can record only horizontal displacements. Two collinear horizontal rods made of fused quartz so as to be insensitive to temperature change are attached to posts about *20* m apart, fixed to the ground; their near ends are separated by a small gap. The changes in separation of the two fixed posts result in changes in the gap width, which are detected with a capacitance or variable-reluctance transducer. In modern instruments the variation in gap width may be observed optically, using the interference between laser light-beams reflected from mirrors attached to the opposite sides of the gap. The strain instrument is capable of resolving strains of the order of 10^{-8} to 10^{-10}.



Schematic design of a strain seismometer (after Press and Siever, 1985).

Equation of the seismometer

Inertial seismometers for recording horizontal and vertical ground motion function on the pendulum principle. When the instrument frame is displaced from its equilibrium position relative to the inertial mass, a restoring force arises that is, to first order, proportional to the displacement. Let the vertical or horizontal displacement, dependent on the type of seismometer, be u and the restoring force $-ku$, and let the corresponding displacement of the ground be q. The total displacement of the inertial mass M is then $u+q$,

The seismogram

A seismogram represents the conversion of the signal from a seismometer into a time record of a seismic event. The commonest method of obtaining a directly visible record, in use since the earliest days of modern seismology, uses a drum that rotates at a constant speed to provide the time axis of the record, as shown. In early instruments a mechanical linkage provided the coupling between sensor and record. The invention of the electromagnetic seismometer by Galitzin allowed transmission of the seismic signal to the recorder as an electrical signal. For many years, a galvanometer was used to convert the electrical signal back to a mechanical form for visual display.

In a galvanometer a small coil is suspended on a fine wire between the poles of a magnet. The current in the coil creates a magnetic field that interacts with the field of the permanent magnet and causes a deflection of the coil. The electrical circuitry of the galvanometer is designed with appropriate damping so that the galvanometer deflection is a faithful record of the seismic signal. The deflection was transferred to a visible record in a variety of ways.

Mechanical and electromagnetic seismometers delivered continuous analog recordings of seismic events. These types of seismometer are now of mainly historic interest, having been largely replaced by broadband seismometers. Galvanometer-based analog recording has been uperseded by digital recording.

Secondary effects of earthquakes: landslides, tsunami, fires and fatalities

Before discussing methods of estimating the size of earthquakes, it is worthwhile to consider some secondary effects that can accompany large earthquakes: landslides, seismic sea waves and conflagrations. These rather exceptional effects cannot be included conveniently in the definition of earthquake *intensity,* because their consequences cannot be easily generalized or quantified. For example, once a major fire has been initiated, other factors not related directly to the size of the earthquake (such as aridity of foliage, combustibility of building materials, availability and efficiency of fire-fighting equipment) determine how it is extinguished.

A major hazard associated with large earthquakes in mountainous areas is the activation of major landslides, which can cause destruction far from the epicenter. In 1970, an earthquake in Peru with magnitude 7.8 and shallow focal depth of 40 km caused widespread damage and a total death toll of about 66,000. High in the Cordillera Blanca mountains above the town of Yungay, about 15 km away, an enormous slide of rock and ice was released by the tremors. Geologists later speculated that a kind of `air cushion' had been trapped under the mass of rock and mud, enabling it to acquire an estimated speed of 300-400 km $h^{-}t$, so that it reached Yungay less than five minutes later. Over 90% of the town was buried under the mud and rock, in places to a depth of 14 m, and 20,000 lives were lost.

When a major earthquake occurs under the ocean, it can actuate a *tsunami* (seismic sea wave). This can be triggered by collapse or uplift of part of the ocean floor, by an underwater landslide, or by submarine volcanism. The tsunami propagates throughout the ocean basin as a wave with period T of around 15-30 min. The velocity of propagation of the wave v is dependent on the water-depth *d,* and the acceleration due to gravity *g,* and is given by: shore of Japan and initiated a tsunami that raced ashore with an estimated wave height of more than 20 m, causing 26,000 fatalities. One of the best-studied tsunami was set off by a great earthquake in the Aleutian islands in 1946. It travelled across the Pacific and several hours later reached Hilo, Hawaii, where it swept ashore and up river estuaries as a wave 7 m high. A consequence of the devastation by this tsunami around the Pacific basin was the formation of the Tsunami Warning System. When a major earthquake is detected that can produce a tsunami, a warning is issued to alert endangered regions to the imminent threat. The system works well far from the source as there is usually adequate warning time, but casualties still occur near to the generating earthquake.

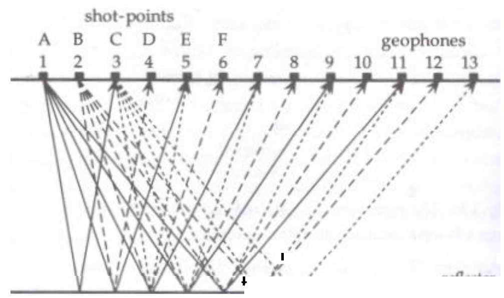
In addition to causing direct damage to man-made structures, an earthquake can disrupt subterranean supply routes (e.g., telephone, electrical and gas lines) which in turn increases the danger of explosion and fire. Aqueducts and underground water pipelines may be broken, with serious consequences for the inhibition or suppression of fires. The San Francisco earthquake of 1906 was very powerful. The initial shock caused widespread damage, including the disruption of water supply lines. But a great fire followed the earthquake, and because the water supply lines were broken by the tremor, it could not be extinguished. The greatest damage in San Francisco resulted from this conflagration.

Earthquake size

There are two methods of describing how large an earthquake is. The *intensity* of the earthquake is a subjective parameter that is based on an assessment of visible effects. It therefore depends on factors other than the actual size of the earthquake. The *magnitude* of an earthquake is determined instrumentally and is a more objective measure of its size, but it says little directly about the seriousness of the ensuing effects. Illogically, it is usually the magnitude that is reported in news coverage of a major earthquake, whereas the intensity is a more appropriate parameter for describing the severity of its effects on mankind and the environment.

Earthquake intensity

Large earthquakes produce alterations to the Earth's natural surface features, or severe damage to man-made structures such as buildings, bridges and dams. Even small earthquakes can result in disproportionate damage to these edifices when inferior constructional methods or materials have been utilized. The intensity of an earthquake at a particular place is classified on the basis of the local character of the visible effects it produces. It depends very much on the acuity of the observer, and is in principle subjective. between rock layers with different seismic velocities and densities are recorded and analyzed. Compactly designed, robust, electromagnetic seismometers - called 'geophones' in industrial usage - are spread in the region of subcritical reflection, within the critical distance from the shot-point, where no refracted arrivals are possible. Within this distance the only signals received are the wave that travels directly from the shot-point to the geophones and the waves reflected at subsurface interfaces. Surface waves are also recorded and constitute an important disturbing `noise', because they interfere with the reflected signal. The closer the geophone array is located to the shot-point, the more nearly the paths of the reflected rays travel vertically. Reflection seismic data are most usually acquired along profiles that cross geological structures as nearly as possible normal to the strike of the structure. The travel-times recorded at the geophones along a profile are plotted as a two-dimensional cross-section of the structure. In recent years, three-dimensional surveying, which covers the entire subsurface, has become more important.



Several field procedures are in common use. They are distinguished by different layouts of the geophones relative to the shot-point. The most routine application of reflection seismology is in *continuous profiling,* in which the geophones are laid out at discrete distances along a profile through the shot-point. To reduce seismic noise, each recording point is represented by a group of interconnected geophones. After each shot the geophone layout and shot-point are moved a predetermined distance along the profile, and the procedure is repeated. Broadly speaking, there are two main variations of this method, depending on whether each reflection point on the reflector is sampled only once *(conventional coverage)* or more than once *(redundant coverage).*

The most common form of conventional coverage is a *split-spread* method, in which the geophones are spread symmetrically on either side of the shot-point. If the reflector is flat-lying, the point of reflection of a ray recorded at any geophone is below the point midway between the shot-point and the geophone. For a shot-point at Q the rays QAP and QBR that are reflected to geophones at P and R represent extreme cases. The two-way travel-time of the ray QAP gives the depth of the reflection point A, which is plotted below the mid-point of QP Similarly, B is plotted below the mid-point of QR. The split-spread layout around the shot-point Q gives the depths of reflection points along AB, which is half the length of the geophone spread PR. The shot-point is now moved to the point R, and the geophones between P and Q are moved to cover the segment RS. From the new shotpoint R the positions of reflection points in the segment BC of the reflector are obtained.

The split-spread method of obtaining continuous subsurface coverage of a seismic reflector.

Common-depth-point method of seismic reflection shooting, showing rays from successive shot-points at A, B and C and the repeated sampling of the same point on the reflector (e.g.,d) by rays from each shot-point. Shotpoint R to the geophone at Q has the same path as the ray QBR from shot-point Q to the geophone at R. By successively moving the shot-point and half of the split-spread geophone layout a continuous coverage of the subsurface reflector is obtained.

Redundant coverage is illustrated by the *common-depthpoint* method, which is routinely employed as a means of reducing noise and enhancing the signal-to-noise ratio. Commonly 24 to 96 groups of geophones feed recorded signals into a multichannel recorder. The principle of common-depth-point coverage is illustrated for a small number of 11 geophone groups in the fig. When a shot is fired at A, the signals received at geophones 3-11 give subsurface coverage of the reflector between points a and e. The shot-point is now moved to B, which coincides with the position occupied by geophone 2 for the first shot, and the geophone array is moved forward correspondingly along the direction of the profile to positions 4-12. From shotpoint B the subsurface coverage of the reflector is between points b and f. The reflector points b to e are common to both sets of data. By repeatedly moving the shot-point and geophone array in the described manner, each reflecting point of the interface is sampled multiply. For example, in the figure, the reflecting point d is sampled multiply by the rays Ad 9, Bd 8, Cd 7, etc. The lengths of these ray paths are different. During subsequent data-processing the reflection travel-times are corrected for *normal moveout,* which is a geometrical effect related to geophone distance from the shot-point. The records are then *stacked,* which is a procedure for enhancing the signal-to-noise ratio.