

ENCYCLOPEDIA ARTICLE

Plate tectonics

Plate tectonics theory provides an explanation for the present-day tectonic behavior of the Earth, particularly the global distribution of mountain building, earthquake activity, and volcanism in a series of linear belts. Numerous other geological phenomena such as lateral variations in surface heat flow, the physiography and geology of ocean basins, and various associations of igneous, metamorphic, and sedimentary rocks can also be logically related by plate tectonics theory.

Theory and Evidence

The theory is based on a simple model of the Earth in which a rigid outer shell 30–90 mi (50–150 km) thick, the lithosphere, consisting of both oceanic and continental crust as well as the upper mantle, is considered to lie above a hotter, weaker semiplastic asthenosphere. The asthenosphere, or low-velocity zone, extends from the base of the lithosphere to a depth of about 400 mi (700 km). The brittle lithosphere is broken into a mosaic of internally rigid plates that move horizontally across the Earth's surface relative to one another. Only a small number of major lithospheric plates exist, which grind and scrape against each other as they move independently like rafts of ice on water. Most dynamic activity such as seismicity, deformation, and the generation of magma occurs only along plate boundaries, and it is on the basis of the global distribution of such tectonic phenomena that plates are delineated. See also: Earth interior; Lithosphere

The plate tectonics model for the Earth is consistent with the occurrence of sea-floor spreading and continental drift. Convincing evidence exists that both these processes have been occurring for at least the last 600 million years (m.y.). This evidence includes the magnetic anomaly patterns of the sea floor, the paucity and youthful age of marine sediment in the ocean basins, the topographic features of the sea floor, and the indications of shifts in the position of continental blocks which can be inferred from paleomagnetic data on paleopole positions, paleontological and paleoclimatological observations, the match-up of continental margin and geological provinces across present-day oceans, and the structural style and rock types found in ancient mountain belts. See also: Continental drift; Continental margin

Plate motion and boundaries

Geological observations, geophysical data, and theoretical considerations support the existence of three fundamentally distinct types of plate boundaries, named and classified on the basis of whether immediately adjacent plates move apart from one another (divergent plate margins), toward one another (convergent plate margins), or slip past one another in a direction parallel to their common boundary (transform plate margins). **Figure 1** shows the major plates of the lithosphere, the major plate margins, and the type of motion between plates. Plate margins are easily recognized because they coincide with zones of seismic and volcanic activity; little or no tectonic activity occurs away from plate margins. The boundaries of plates can, but need not, coincide with the contact between continental and oceanic crust. The nature of the crustal material capping a plate at its boundary may control the specific processes occurring there, particularly along convergent plate margins, but in general plate tectonics theory considers the continental crustal blocks as passive passengers riding on the upper surface of fragmenting, diverging, and colliding plates.

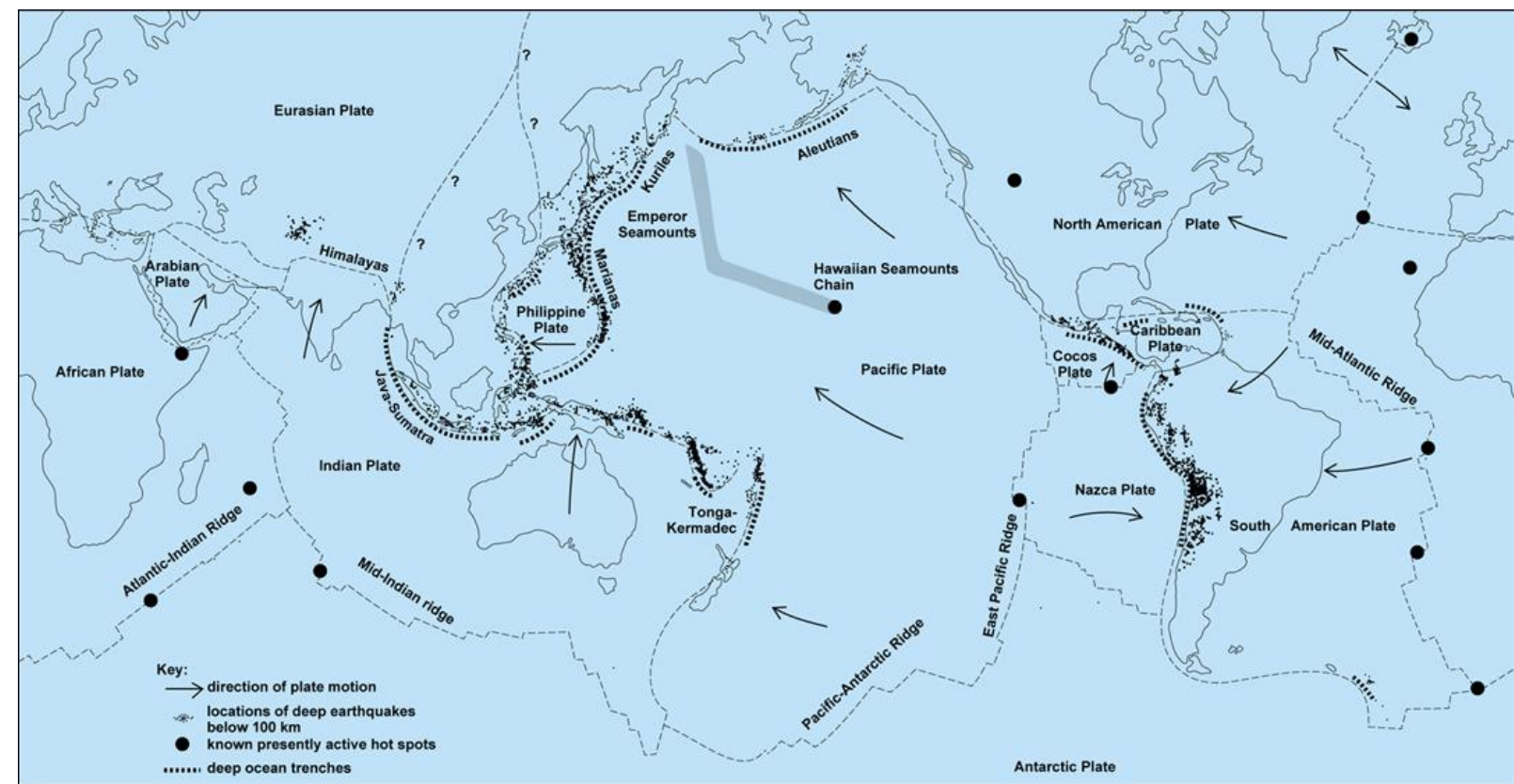


Fig. 1 Tectonic map of the Earth. (After M. Nafi Toksoz, *The subduction of the lithosphere*, *Sci. Amer.*, 233(5):88–98, November 1975)

The velocity at which plates move varies from plate to plate and within portions of the same plate, ranging between 0.8 and 8 in. (2 and 20 cm) per year. This rate is inferred from estimates for variations in the age of the sea floor as a function of distance from mid-oceanic ridge crests. Ocean-floor ages can be directly measured by using paleontological data or radiometric age-dating methods from borehole material, or can be inferred by identifying and correlating the magnetic anomaly belt with the paleomagnetic time scale.

Divergent plate margins

As the plates move apart from the axis of the mid-oceanic ridge system, the new volcanic material welling up into the void forms a ribbon of new material, which gradually splits down its center as the boundary of plate separation continues to develop. Each of the separating plates thus accretes one-half a ribbon of new lithosphere, and in this way new lithosphere and hence new surface area are added. The process is considered to be continuous, and the boundary at which separation is taking place always maintains itself in the center of the new material. See also: Mid-Oceanic Ridge

The accretion at any spreading boundary is usually bilaterally symmetric. The morphology of the ridges is also quite symmetric and systematic. The new material that wells up at the ridge axis is hot and therefore expanded and less dense than the surrounding older material. Consequently, the new material is topographically highest. As new material divides and moves away from the ridge axis, it cools, contracts, becomes denser, and subsides. The densification is caused by the combined effect of pure thermal contraction and thermally driven phase changes. Subsidence is fastest for newly generated oceanic crust and gradually decreases exponentially with time. This observation explains the fact that in cross section the shape of the slope of the ridges is steepest at the ridge axis and gradually decreases down the flanks beneath the abyssal sediments and to the bounding continents. Since all known oceanic lithosphere has been generated by the spreading at a ridge axis, all oceanic lithosphere is part of the mid-oceanic ridge system. Because of the systematic way in which the morphology of the ridges is formed, most oceanic crust follows the same time-dependent subsidence curve within an error range of about 300 ft (100 m). This means that the same age-versus-depth curve fits nearly all parts of the mid-oceanic ridge system. The ridge axis is found at a depth of 1.71 mi \pm 300 ft (2.75 km \pm 100 m), and oceanic lithosphere that is 30 m.y. old is found at a depth of 2.72 \pm 300 ft (4.37 km \pm 100 m). There are exceptional areas such as Iceland where the ridge axis is above sea level.

Magnetic lineations and age of oceans

Much of the evidence which leads to the development of the concept of sea-floor spreading, and in turn to an understanding of divergent plate margins and plate tectonics theory, came from analyses of the magnetic properties of the sea floor. Magnetometer surveys across the sea floor near mid-oceanic ridges reveal a pattern of alternating positive and negative magnetic anomalies (Fig. 2). The characteristics of these magnetic anomaly patterns (parallel to and in symmetrically matching widths across the ridge crest) and present-day general knowledge of the Earth's magnetism logically support the conclusion that sea-floor spreading occurs.

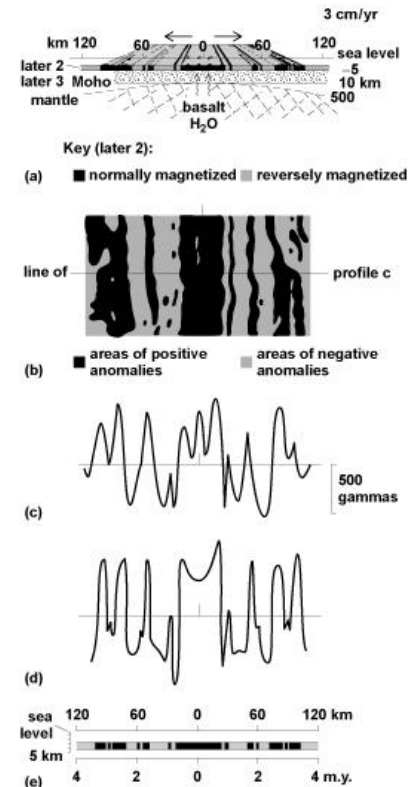


Fig. 2 Tape-recorder conveyor-belt scheme for sea-floor spreading. (a) Schematic of crustal model, applied to Juan de Fuca Ridge, southwest of Vancouver Island. (b) Part of summary map of magnetic anomalies recorded over Juan de Fuca Ridge. (c) Total-field magnetic anomaly profile along the line indicated in (b). (d) Computed profile assuming the model and reversal time scale. Intensity and dip of Earth's magnetic field taken as 54,000 gammas (0.54 oersted) and $+667^\circ$; magnetic bearing of profile 087° ; 10 km horizontally is equivalent to 100 gammas vertically. Normal or reverse magnetization is with respect to an axial dipole vector, and the assumed effective susceptibility is 0.01 except for central block and ridge crest ($+0.02$). (e) Scale applicable to (b)-(d). 1 cm = 0.4 in.; 1 km = 0.6 mi; 1 gamma = 1 nanotesla. (After R. A. Phinney, ed., *The History of the Earth's Crust*, 1968)

The Earth's magnetic field as measured at the surface may be approximately represented by an axial geocentric dipole—in other words, as if a simple two-pole magnet existed coincident with the Earth's rotational axis. The Earth's magnetic field reverses polarity episodically. Over an interval of time, the strength of the dipole field gradually decreases to zero and then gradually increases in the opposite direction. The total transit time from full strength in one direction (polarity) to full strength in the opposite direction may be less than 6000 years. The residence time in any one particular polarity ranges from 10,000 years up to tens of millions of years. Thus, the polarity behavior of the Earth's magnetic field is approximately analogous to a randomly triggered flip-flop circuit.

The top 600 to 1500 ft (200 to 500 m) of the lithosphere is the oceanic basalt, which is accreted at separating boundaries and which contains magnetizable iron minerals. As the newly injected material cools through its Curie point (-1100°F or 570°C), it becomes permanently magnetized in the direction of the Earth's magnetic field at that time and place. Because the oceanic crust is continuously being formed and transported in a bilaterally symmetric pattern away from the line of rifting, it is similar to a continuous magnetic tape record of the Earth's magnetic polarity events. At the mid-oceanic ridge axis, the present polarity of the field is recorded. Down the ridge flanks, progressively older oceanic crust is traversed; hence the crust contains in its remanent magnetism a continuously older record of the polarity of the field. See also: Basalt; Curie temperature

As a result of sea-floor spreading and polar reversals, the oceanic crust appears magnetically as a set of alternately normally or reversely magnetized strips of basalt arranged in a bilaterally symmetrical pattern about the mid-oceanic ridge axes. Because the rate of separation at most segments of the mid-oceanic ridge axes has varied only slowly with time, the spatial distribution of the stripes is proportional to the temporal history of the polarity of the field. Thus, the polarity of the oceanic crust may be inferred indirectly from the pattern of magnetic anomalies.

Because the occurrence of polarity events is random, they form a unique sequence which is reflected directly in the magnetic anomaly patterns. A unique pattern of magnetic anomalies has been correlated throughout most of the world's oceans. From this pattern a magnetic reversal time scale from the present to 160 m.y. ago has been developed. This geomagnetic time scale has been calibrated by drilling on key anomalies in the oceans and paleontologically determining the age of the sediment-basalt interface. Thus a magnetic polarity time scale has been developed, and with it identifiable magnetic anomalies that can be correlated with part of the time scale may be assigned an absolute age. In this way, the spatial distribution of the magnetic anomalies is used to calculate the rates of separation of the plates in the past. See also: Geomagnetism; Paleomagnetism

Convergent (destructive) plate margins

Because the Earth is neither expanding nor contracting, the increase in lithosphere created along divergent boundaries must be compensated for by the destruction of lithosphere

elsewhere. The rates of global lithosphere construction and destruction must be equal, or the radius of the Earth would change. Compensatory destruction or removal of lithosphere occurs along convergent plate margins (subduction zones) and is accomplished by plate subduction and continental collision. See also: Subduction zones

Along subduction zones, one plate plunges beneath another (**Fig. 3**). The downgoing slab is usually oceanic because the relatively buoyant continental lithosphere cannot be subducted beneath the relatively denser oceanic lithosphere. The upper or overriding plate may be continental or island arc lithosphere, and occasionally oceanic plateaus as in the case of Oregon and Alaska.

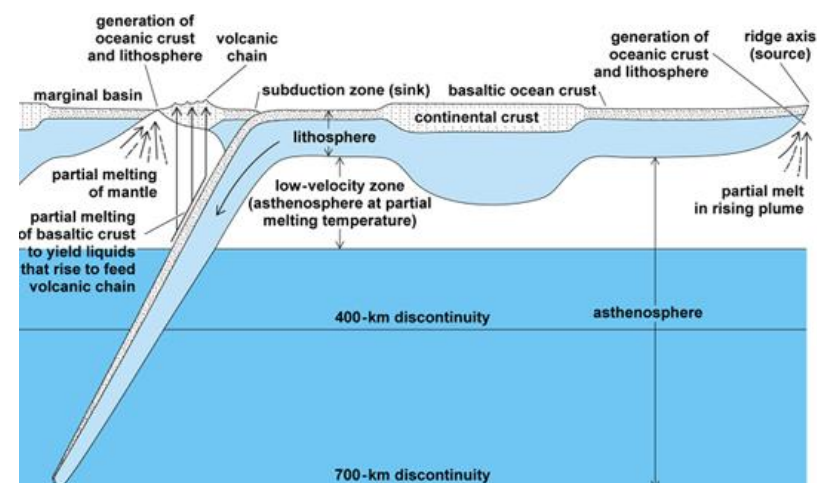


Fig. 3 Cross section of the upper mantle and crust showing a lithospheric plate riding on the asthenosphere. The continent is embedded in the plate and moves with it. Divergent plate margin with generation of oceanic crust and lithosphere is shown on the right; convergent plate margin with subduction is shown on the left. 1 km = 0.6 mi. (After J. F. Dewey, *Plate tectonics*, *Sci. Amer.*, 226:56–68, May 1972)

The dip of the downgoing underthrust slabs varies but averages 45°. Earthquake foci along individual subduction zones indicate the plate underthrust angle, often referred to as the Benioff plane or Wadati-Benioff zone. This plane dips away from oceanic trenches toward adjacent volcanic arcs and continents, and marks the surface of slippage between the overriding and descending lithospheric plates. Successive belts of shallow (less than 40 mi or 70 km), intermediate (40–200 mi or 70–300 km), and deep-focus (200–400 mi or 300–700 km) earthquakes are associated with subduction zones. Subduction zones are also associated with active volcanism and the development of deep-ocean trenches. These features encircle the Pacific Ocean basin.

A line of andesitic volcanoes usually occurs on the upper plate, forming a chain that is parallel to the trench. The volcanism occurs at that point above the subduction zone where the upper surface of the downgoing lithosphere has reached a vertical depth of approximately 75 mi (120 km). The volcanic materials arise from partial melting of the downgoing slab. The volcanoes of the Andes chain of South America and island arcs in the Pacific such as Izu-Bonin and the Mariana have formed in this manner.

Convergent plate margins that develop island arc systems adjacent to trench systems are geologically complex. The region between a volcanic island arc and the trench consists of volcanic rock and sediments. Slumping and turbidity currents transport sediment into the trench axis. The trench sediment may be transferred from the downgoing plate to the continent or island arc. Conversely, a convergent margin without a large sediment supply may be eroded during subduction.

Because of density consideration, subduction requires at least one of the two converging plates to be oceanic. If both converging plates consist of continental lithosphere, continental collision occurs, but not subduction. Continental collision results in compensatory reduction in lithospheric width by folding and compressing the lithosphere into narrower, linear mobile belts. In such collisions, sediments deposited along the continental margins and within the closing ocean basins are compressed into a series of tight folds and thrusts. Fragments of oceanic crust may be thrust up onto adjacent continental rocks (obduction) as ophiolite successions. A classic example of a continental collision belt is the Himalayan belt, produced during the Cenozoic Era by the convergence of the Indian continent with Eurasia.

Like plate divergence, plate convergence produces a distinctive suite of igneous rock types. Subduction zones are marked by the belts of predominantly andesitic volcanoes either in island arcs located landward of the trench system (Japan and the Philippines) or along the rim of overriding continental blocks (the Andes belt). These andesitic volcanic terranes are commonly associated intimately with plutonic igneous rocks, mainly granodiorites. The origin of andesitic magmas, and the predominance of granodiorite plutons within continental blocks to the exclusion of most other igneous rock varieties, was perplexing until the development of plate tectonics theory. Now both seem to be directly related to the generation of parent magmas by the frictional melting of ocean-floor basalt and overlying sediment cover along subduction zones. Partial melting of the lower crust and upper mantle also occurs.

Plate subduction and continental collision can also explain the origin of two other puzzling rock sequences commonly found within mountain belts: mélangé and blueschist terranes. Mélangé, a heterogeneous assemblage of intensely sheared, poorly sorted, angular blocks set in a fine-grained matrix, is probably generated at shallow depths along subduction zones as the oceanic crust and overlying sediment cover of the descending plate are scraped and crushed against the overriding plate. Blueschist terranes (their dark blue color is imparted by the presence of various low-temperature–high-pressure metamorphic minerals such as glaucophane, lawsonite, and jadeite) occur in belts within mountain chains, parallel with but external to (toward the ocean) the more conventional paired metamorphic facies of the greenschist terrane. The peculiar physical conditions required by the blueschist facies, great burial depth (in excess of 10 mi or 16 km) but moderate temperature (390–800°F or 200–450°C), should be generated along subduction zones when the descending lithospheric plate is underthrust at a greater rate than the local geothermal gradient can heat it. See also: Blueschist

Transform plate margins

Transform faults are always strike-slip faults. They occur where the relative motion between the two plates is parallel to the boundary that separates the plates. They may join a ridge to a ridge, a ridge to a trench, or a trench to a trench. Ridge-trench transforms will always change length with time. A trench-trench transform may lengthen, shrink, or remain constant, depending on which of the plates that form the subduction system is the downgoing plate. A transform that joins two ridge axes will not change in length with time.

First, consider ridge-ridge transforms (**Fig. 4**). Earthquake epicenter data show that earthquakes occur only along the ridge axes and the connecting transform. Studies of earthquakes that occur at a ridge-ridge transform fault show that the first motion was strike-slip parallel to the direction of the transform and opposite to the sense of offset of the ridge axis. If the ridge axes are offset left-laterally, the relative motion across the transform that joins the ridges will be right-lateral as the plates separate. The motion of the fault takes place spasmodically. Deformation occurs on either side of the fault until the elastic limit is reached and the rupture occurs, causing an earthquake. See also: Earthquake

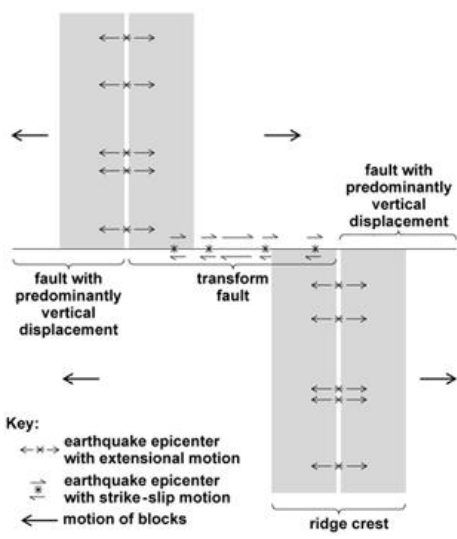


Fig. 4 Idealized map view of a transform fault. The ridge appears to be offset in a left-lateral sense, but the motion on the transform fault between the ridge crests is right-lateral. (After C. K. Seyfert and L. A. Sirkin, *Earth History and Plate Tectonics*, Harper Row, 1973)

Fracture zones are morphologic scars that are fossilized transform faults. They may be used to determine the past direction of relative motion of the bounding plates. In **Figs. 5** and **6**, the transform fault A-B offsets two ridge axes, and there is right-lateral strike-slip motion across the transform. All the material to the left of the ridge-transform-ridge boundary A'ABB' belongs to the torsionally rigid plate I, and all the material to the right of this boundary is part of plate II. Thus, beyond the ends of the transform, to the left of A' and right of B, there is no relative horizontal motion.

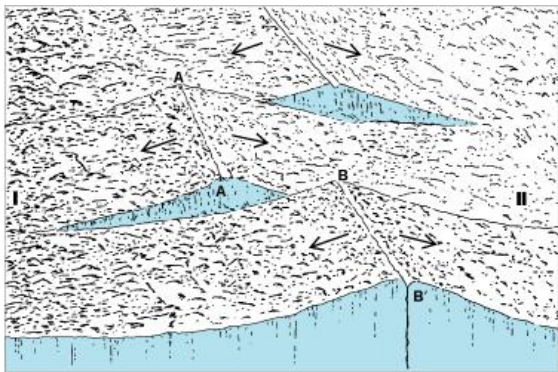


Fig. 5 Ridge-ridge transform fault appears between two segments of ridge that are displaced from each other. (After J. Tuzo Wilson, ed., *Continents Adrift and Continents Aground*, W. H. Freeman, 1976)

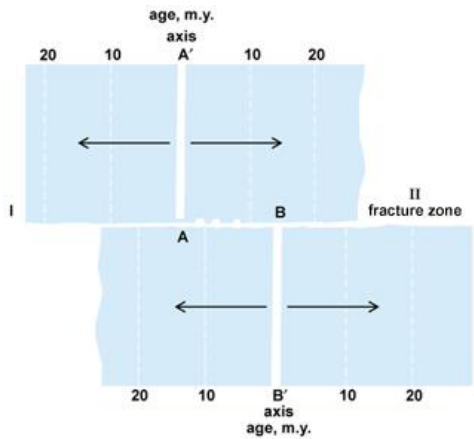


Fig. 6 Molten rock wells up from the deep Earth along a spreading axis, solidifies, and is moved out (shown by arrows). The axis is offset by a transform fault. Between two offset axes, material on each side of the transform fault moves in opposite directions, causing shallow earthquakes. (After J. R. Heirtzler, *Sea-floor spreading*, *Sci. Amer.*, vol. 219, no. 6, December 1968)

Because of the way in which fracture zones are formed, they also preserve a record of the past direction of relative motion of the plates. The transform fault between plates I and II in **Figs. 5** and **6** is in theory a vertical plane between these plates along which horizontal strike-slip motion occurs. The face of plate I that abuts against plate II in effect defined this transform plane. But as plate I moves to the left, this transform plane or face passes by the end of ridge A'A at point A, and new younger material is intruded against it. Together, the transform plane of plate I and the new younger material move off to the left. This process is continuous. The new younger material is more elevated and forms an escarpment whose horizontal trace records the direction of relative movement. Because the rate of subsidence decreases with time, the difference in elevation across a fracture zone decreases with age.

A very narrow and deep rift valley is also associated with the transform. However, because of time-dependent cooling, the oceanic lithosphere contracts horizontally. This contraction causes cracking of the lithospheric plate and widening and splaying of the rift associated with the fracture zone. Young fracture zones are characterized by an escarpment and a deep fissure along their length. The magnitude of the vertical offset decreases with age, and the fissure tends to widen. Down the lower flanks of the ridges, turbidities from the abyssal plains often finger into the fracture zone. The fracture zones form long curvilinear features throughout the oceans and are mappable because of morphology alone. See also: Rift valley

The numerous strike-slip faults, which offset segments of the mid-oceanic ridge system, are classical examples of transform plate boundaries. The San Andreas system of California, which offsets portions of the East Pacific Rise, is probably the best-known example. Transform faults show apparent lateral displacements of many tens or even hundreds of kilometers. Occasionally extension, and consequent igneous activity, may occur at a transform fault.

Transform faults are parallel to the direction of relative motion between the bounding plates. According to a theorem of spherical geometry, if plate 1 is moving with respect to plate 2 on a sphere, the instantaneous relative motion may be represented by the rotation of one plate with respect to the other about a single stationary pole. Circles drawn on the sphere concentric to the pole of relative motion will be parallel to the direction of instantaneous relative motion. Great circles drawn perpendicular to these concentric circles will intersect at the pole of instantaneous relative motion. The transform faults that separate block 1 from block 2 are parallel to the direction of present-day relative motion and therefore must lie on circles that are concentric to a pole of relative motion of 1 with respect to 2. Great circles drawn perpendicular to the transform faults will intersect at the pole of relative motion. In this way, the pole of relative motion may be found for any two plates whose common boundary consists in part of transform faults.

The instantaneous rate of relative motion is an angular rate of rotation of 1 with respect to 2 about an axis through the pole of relative motion and the center of the Earth. The linear rate of relative motion R (in centimeters per year) will then be proportional to the sine of the colatitude with respect to the instantaneous pole,

$$R = a \sin \theta \frac{d\phi}{dt}$$

as shown in the equation below, where a = the radius of the Earth, 2.4×10^8 in. (6×10^8 cm); θ = the colatitude of the point at which the rate is to be determined with respect to the instantaneous pole; and $d\phi/dt$ = the angular rate of motion of 1 with respect to 2.

In this way, spreading-rate data calculated from the distribution of magnetic lineations near the axis of a spreading ridge-transform system may be used to compute a pole of relative motion. Note that in this case the pole is not strictly instantaneous because magnetic lineations covering a finite period must be used. It has been found that the poles and rates of relative motion tend to remain nearly constant, perhaps changing quite slowly over long time intervals. Thus, poles and rates computed in the above fashion may often be used to describe the relative motion for several tens of millions of years.

The fracture zones may often be treated as fossil transform faults. To obtain paleopoles of relative motion, segments of fracture zones spanning the same age and from the same plate must be used. Perpendicular great circles drawn to these segments give a pole of relative motion for that time interval in a reference frame fixed with respect to that plate. The magnetic anomaly lineations abutting the fracture zone are used to determine the rate of motion. By calculating a time sequence of poles and rates using time sequential segments of fracture zones, the spreading history of an entire ocean basin may be determined. See also: Fault and fault structures; Transform fault

Triple junctions

The point where the boundaries between three pairs of plates join is known as a triple junction. Because the Earth's lithosphere shell is segmented into a mosaic, the boundary between any two plates must end in a triple junction. Quadruple junctions are theoretically possible, but except under very unusual circumstances they degenerate immediately into two triple junctions. Figure 1 shows a number of triple junctions that vary as to the types of boundaries that meet and as to the geometry.

When three bodies are moving with respect to each other across a spherical surface, if the relative motion of each of two pairs of the plates is known, the relative motion of the third pair may be determined. In the case of a spherical surface, the relative motion between two plates may be represented by a vector through the pole of instantaneous relative motion, with the length of the vector proportional to the rate of rotation. The relative motion of three plates or bodies on a sphere may be represented by three such vectors. Each vector gives the instantaneous relative motion of two of the three plates. Thus, three such vectors describe completely the relative motion of the three plates and must add to zero.

The orientation of the boundary between these plates will determine the type of interaction at the boundary (that is, spreading, subduction, or transform). There are a number of possible triple junctions. The Cocos, Pacific, and Nazca plates join at a ridge-ridge-ridge triple junction (Fig. 1). This is an example of a stable triple junction, that is, the triple junction moves in a constant direction and at a constant speed with respect to each of the plates. Other triple junctions are unstable or transitional, such as a ridge-ridge-transform triple junction which will change immediately to a transform-ridge-transform triple junction. The most geologically important triple junctions are those that migrate along a boundary. Consequently, the boundary along which the triple junction moves will experience a change in the direction of relative motion. If the change in relative motion is radical, there may be an accompanying change in tectonic style. This type of plate interaction occurred at the western boundary of the North American Plate during the Cenozoic, and has resulted in a time transgressive change, from south to north, from subduction to strike slip. This is reflected in the geologic record as a change from compressive tectonics with thrusting, folding, and Andean or island arc volcanism to strike slip, as occurs along the San Andreas Fault.

Paleogeography

Because the accretion of new lithosphere is bilaterally symmetric, the associated magnetic lineations must also be bilaterally symmetric across a ridge axis. Once formed, oceanic crust is rarely disturbed by horizontal shear, but each magnetic lineation preserves the shape of the mid-oceanic ridge axis and offsetting transforms at the time of its formation. The fact that lineations of the same age from opposite sides of a ridge axis may be fitted together is considered evidence of the undisturbed nature of the oceanic crust.

The North and South Atlantic oceans were formed by the rifting and drifting of the continents that now surround the Atlantic. These continents have been rifted as part of the separating lithospheric plates since the Triassic as the Atlantic Ocean slowly opened. The magnetic lineation pattern in the North Atlantic records this history of separation of Africa, Europe, and North America. When magnetic lineations of the same age from opposite sides of the ridge axis are fitted together, the relative positions of the continents at the time the lineations formed are obtained. This is equivalent to reversing the process of sea-floor spreading. In effect, the material younger than the two lineations is removed, and the older portions of the plates with the continents move toward each other until the two lineations fit together.

This technique is applicable to any ocean where separation at ridge axes has led to the passive rifting of the surrounding continents, such as the southeast and central Indian Ocean, the South and North Atlantic, the Labrador and Norwegian seas, and the Arctic Ocean.

The above technique gives the position of the continental landmasses relative to each other, not their position relative to the rotational axis or paleomagnetic data. It is assumed that the Earth's magnetic field, now approximated by an axial geocentric dipole, has been similarly oriented throughout geologic time. Subaerial volcanic rocks acquire a remanent magnetism in the same way as oceanic crust, by cooling below the Curie point. Subaerial sedimentary rocks may also be remanently magnetized by the process of detrital remanent magnetization. As iron-bearing particles settle from moving currents, their orientation is influenced by the prevailing magnetic field. By measuring the direction of the remanent magnetism of both igneous and sedimentary rocks whose ages can be determined, the position of the paleomagnetic pole for that age can be found. From these data, the paleolatitude may be found for the set of continental landmasses whose positions relative to each other are known (**Fig. 7**). See also: Paleogeography; Rock magnetism

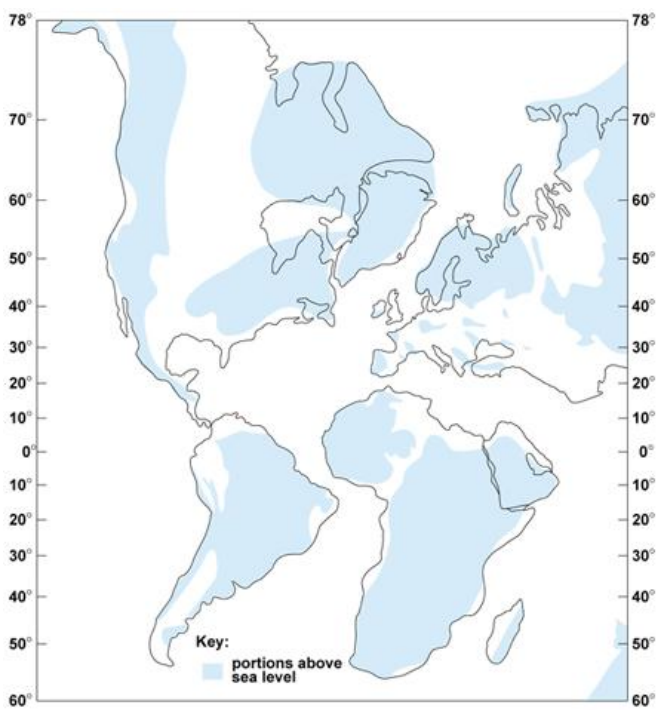


Fig. 7 Map illustrating the paleogeographic arrangement of the Atlantic continents during Coniacian time. Sea level may have been over 300 m (1000 ft) above present level at that time because of the volume of the mid-oceanic ridge system. Shaded areas are portions above sea level.

Sea-level changes

Many important geologic phenomena may in part be related to plate tectonics processes, for example, changes in sea level and climate and changing patterns of evolution. The geologic record contains evidence of almost continuous eustatic sea-level changes throughout the Phanerozoic. Several large and rapid sea-level changes can be attributed to fluctuations in the volume of massive continental glaciers. However, it is apparent that large sea-level changes (greater than 300 ft or 100 m) also occurred during periods when continental glaciers were small or nonexistent.

Several factors affect the volume of the ocean basins or the ocean waters and can thus alter eustatic sea level. One of the most important of these in terms of magnitude and rate of sea-level change is variation in the volume of the mid-oceanic ridge system. As explained above, once formed, the sea floor subsides systematically with time, with all oceanic crust following the same exponential subsidence curve. The volume of the mid-oceanic ridge system is quite large. If sea-floor spreading were to cease today, 70 m.y. from now the ridges would have subsided sufficiently to decrease the depth of the water over the abyssal area by about 1500 ft (450 m). The continental freeboard would increase by 1100 ft (320 m).

The volume of the mid-oceanic ridge system may be altered in several ways, one of which is to change the spreading rate. Because all ridges follow the same subsidence curve (which is a function of time only), age-versus-depth relationships are the same for all ridges. This means that if two ridges have been spreading at a constant but different rate for 70 m.y., the ratio of their volumes per unit length will equal the ratio of their spreading rates. If the spreading rate of the faster ridge is reduced to that of the slower ridge, volume/unit length of the larger ridge will gradually be reduced to that of the slower ridge. If spreading rates decrease, ridge volume also decreases, increasing the freeboard of the continents. The converse is also true: by increasing spreading rates, ridge volume will increase and the freeboard of the continents will be reduced. Several other ways of changing the volume of the mid-oceanic ridge system exist: ridges may be destroyed; segments of the ridge system may be subducted; and new ridges may be created by continental rifting and new rifting within ocean basins.

Both the length of the mid-oceanic ridge system and the spreading rate at its various segments have varied considerably during Phanerozoic time. These changes have caused large variations in sea level. For example, by using magnetic anomaly data to calculate spreading rates and ridge lengths back to the Upper Cretaceous, it has been estimated that the sea level then may have been as much as 900 ft (300 m) above present (Fig. 8).

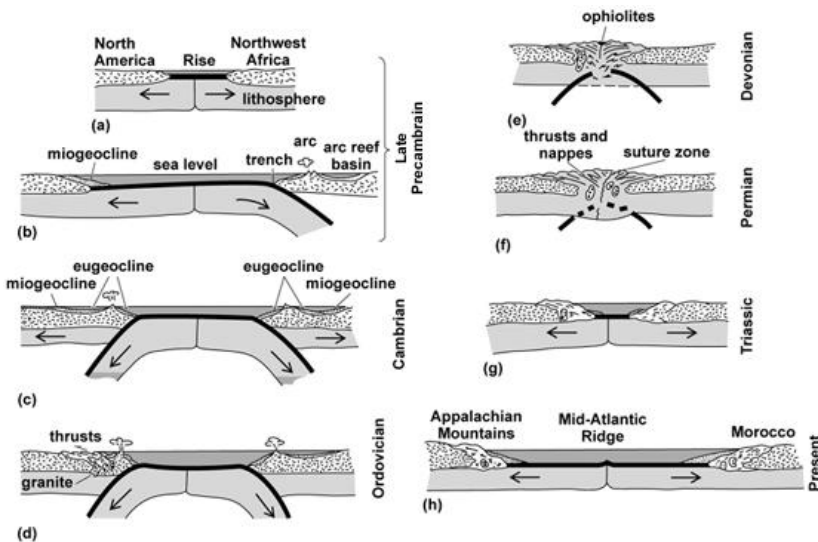


Fig. 8 Schematic representation of (a-h) the sequence of changes in the mid-oceanic ridge system in the plate tectonics history of the Appalachian orogenic belt. (After K. C. Condie, *Plate Tectonics and Crustal Evolution*, Pergamon, 1976)

Hot spots and mantle plumes

The existence of convective plumes originating in the deep mantle from below the level of the asthenosphere and rising to the bottom of the lithosphere has been proposed. About 14 major convective plumes are believed to exist today (Fig. 1). These plumes are believed to be nearly stationary with respect to each other, and hence they may be used as reference points with respect to which all the plates are moving. Mantle plumes, rather than convection currents, may also be the driving mechanism for plate motion. The latter two hypotheses are not necessary conditions for the validity of the first. In other words, there may be mantle plumes that rise to the bottom of the lithosphere, but they may not be stationary with respect to each other and may not furnish the plate driving mechanism. The major clue to the existence of hot spots is the lines of intraplate volcanoes that are left as a trace of the passage.

As a plate passes over a hot spot, the hot spot burns its way through the plate. This releases volatile and eruptive magma on the plate surface. A classic example of the “volcanic” consequences of passage of a plate over a hot spot is the Hawaiian-Emperor Seamount chain. Hawaii and the islands immediate to it are volcanically active at present. The rest of the islands and seamounts of the Hawaiian and Emperor chains are inactive, but all are of volcanic origin. The islands and seamounts are found to be sequentially older west-northwestward along the Hawaiian chains and are progressively older north-northwestward along the Emperor Seamount chain. As an explanation, it has been hypothesized that the Pacific Plate moved first north-northwesterly and then west-northwesterly over a single hot spot. Other seamount chains in the Pacific, such as the Line Islands, the Tuamotu Archipelago, and the Austral Seamount chain, are presumed to be of similar origin. See also: Hot spots (geology); Magma; Oceanic islands

In the Atlantic, features that are regarded as major hot-spot traces are Iceland and the Iceland Faroes Ridge, the New England Seamount chain, the Columbia Seamount chain, the Rio Grande Rise, and Walvis Ridge; and in the Indian Ocean, the Ninety-East Ridge is an analogous feature. The age progression along each of these features has been estimated; the theory that these hot spots have remained fixed in position relative to each other has been tested. The results indicate that some relative motion may occur, but it appears to be an order of magnitude less than the generally observed rate of interplate motion.

The hot spot hypothesis is a significant complement to plate tectonics. There is little doubt that hot spots or plumes of some type do occur. Intraplate volcanism often seems to be a manifestation of their existence. If, in addition, the hot spots prove to be essentially stable with respect to the rotational axis, they may be useful as a reference frame. They could serve as a latitudinal constraint in addition to paleomagnetic data. They could also be very useful in finding the relative motion between plates that have been separated by subduction zones or transform faults (that is, Pacific–North America relative motion).

Earth history

Not only does plate tectonics theory explain the present-day distribution of seismic and volcanic activity around the globe and physiographic features of the ocean basins such as trenches and mid-oceanic rises, but most Mesozoic and Cenozoic mountain belts appear to be related to the convergence of lithospheric plates. Two different varieties of modern mobile belts have been recognized: cordilleran type and collision type. The Cordilleran range, which forms the western rim of North and South America (the Rocky Mountains, Pacific Coast ranges, and the Andes), has for the most part been created by the underthrusting of an ocean lithospheric plate beneath a continental plate. Underthrusting along the Pacific margin of South America is causing the continued formation of the Andes. The Alpine-Himalayan belt, formed where the collision of continental blocks buckled intervening volcanic belts and sedimentary strata into tight folds and faults, is an analog of the present tectonic situation in the Mediterranean, where the collision of Africa and Europe has begun. See also: Cordilleran belt; Mountain systems; Orogeny

Abundant evidence suggests that sea-floor spreading, continental drift, and plate tectonics have occurred for at least the past 600 m.y., during Phanerozoic time. Furthermore, it is probable that plate tectonics phenomena have dominated geologic processes for at least 2.5 billion years (b.y.). Present-day ocean basins are young (post-Paleozoic) features, and their origin is adequately explained by sea-floor spreading. The origin and evolution of the continental blocks, substantial portions of which are Precambrian in age, probably may be explained by accretion at subduction zones. Continental blocks, though largely composed of granite and granodiorite, appear to be mosaics of ancient mobile belts that progressively accreted laterally, vertically, or both, through time. Many late Precambrian and Paleozoic mobile belts, such as the Appalachian-Caledonian belt of the North Atlantic region, are interpretable in terms of plate tectonics mechanisms (Fig. 8). Initial rifting breaks a preexisting supercontinental landmass into two or more fragments that drift apart at the flanking margins of the growing ocean basin. At the margin of the growing basin, continental shelves, continental rises, and abyssal plains accumulate sediments. Cessation of sea-floor spreading and the development of new convergent margins along one or both sides of the ocean basin lead first to cordilleran-type orogeny and subsequently to continental collision. These orogenies suture or stitch together the original continental fragments and laterally buckle the intervening sediments and volcanic belts into folds and faults, initiate metamorphism, and generate volcanic and plutonic magmas. Belts may later be bisected by the development of new divergent plate margins that produce ocean basins whose axes may cut across the older mountain chains. This seems to be the case with the Appalachian-Caledonian belt, where the Mesozoic-Cenozoic development of the Atlantic Ocean has separated the chain into the Appalachians in North America and the Caledonides in Greenland, Great Britain, and Scandinavia. Most Paleozoic mobile belts, and many Precambrian belts at least 2.5 b.y. old, can be interpreted in an analogous fashion, for they contain the various petrogenetic associations of ocean opening and suture: ophiolites, blueschist terranes, mélanges, and a suite of sedimentary rocks whose composition, texture, and distribution suggest deposition in trenches and inland seas, on the abyssal plains of ocean basins, and along continental shelves and rises. Because the Earth itself has evolved, chemical and petrologic aspects of these more ancient mobile belts often differ in detail from more modern analogs. Their organization leaves little doubt that they were formed by plate tectonics processes.

Plate tectonics is considered to have been operative as far back as 2.5 b.y. Prior to that interval, evidence suggests that the plate tectonics may have occurred, although in a markedly different manner, with higher rates of global heat flow producing smaller convective cells or more densely distributed mantle plumes which fragmented the Earth's surface into numerous small, rapidly moving plates. Repetitious collision of plates may have accreted continental blocks by welding primitive “greenstone island arc terranes” to small granitic subcontinental masses. See also: Continents, evolution of; Geodynamics

Walter C. Pitman III

Details of Processes at Convergent Margins

Convergent continental and island arc margins encircle the Pacific Ocean and occur locally in all other oceans (**Fig. 9**). At convergent margins, two of the Earth's lithospheric plates are directed against each other. Normally, one plate consists of dense oceanic lithosphere which underthrusts, or subducts, beneath less dense continental or island arc lithosphere (**Fig. 10**).

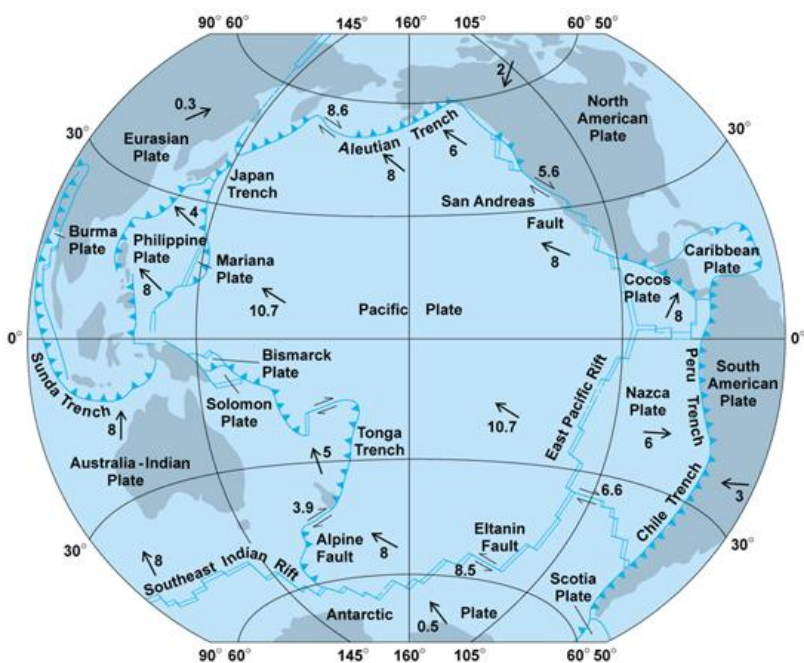


Fig. 9 Plate tectonic map of the circum-Pacific region showing major zones of plate convergence and spreading ridges. Arrows and numbers indicate the direction and the rate (cm/yr; 1 cm = 0.4 in.) of plate movement. The thrusts of the zones of convergence are marked with teeth on the upper plate. (After G. W. Moore, *Plate-Tectonic Map of the Circum-Pacific Region: Explanatory Notes*, American Association of Petroleum Geologists, 1982)

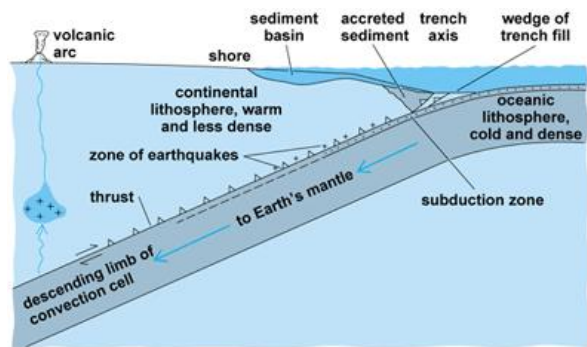


Fig. 10 Cross section through a typical convergent margin. The relatively dense oceanic lithosphere sinks beneath the relatively light continental lithosphere and forms the descending limb of a convection cell in the Earth's mantle.

Structure and tectonic units

The thrust fault between the plates is the major tectonic feature of a subduction zone, and opposing plate motion causes great earthquakes that begin at 10–20 km (6–12 mi) depths in the seismogenic zone. As the plates shear past one another, sediment and rock of the subducting plate may be added to the upper plate and upper-plate material may be eroded, introducing terrigenous material into the subduction zone. The dominant material transfer is a basis for dividing convergent margins into accretionary and erosive types. In the absence of a standard nomenclature, the following is used here. The upper plate of convergent margins is commonly formed by four units, consisting of a frontal prism that is sometimes backed against an accreted prism, which in turn backs against the core rock framework and is overlain by a slope apron. The lower plate consists of igneous rock overlain by oceanic and trench fill sediment. Sediment that stays attached to the lower plate when introduced into the subduction zone forms the subduction channel that is capped by the interplate thrust. See also: Marine geology

The frontal prism begins at the initial contact between converging plates as represented by the first fault that displaces the trench axis sea floor, referred to as the deformation front. Rapid deformation characterizes the frontal prism. The prism extends to either a more rigid core framework or a well-consolidated accreted prism.

Compressional deformation tectonically thickens the prism in contrast to less prominent active tectonism landward. The frontal prism consists of material derived from off-scraped trench fill, slope apron sediment, or fragmented basement rock that was detached at the basement–prism contact. Along more than 50% of all convergent margins where trench fill is no thicker than 1 km (0.6 mi), frontal prisms are less than 20 km (12 mi) wide. Once they reach 25–30 km (15–20 mi) in width, material at the back of the prism has consolidated, and it becomes sufficiently rigid to become part of the accreted prism. Thus, the contact between frontal prism and accreted wedge migrates toward the trench axis. Frontal prisms widen continents only a small amount in contrast to accreted prisms.

The accreted prism, 30 to more than 100 km (20 to more than 60 mi) wide, is a body of oceanic and trench fill sediment that was transferred from the subducting lower plate at the frontal prism and became attached to the upper plate framework. Accreted prisms develop from the frontal prism and consist of consolidated and rigid sedimentary rock with porosities of about 15% or less. The accreted prism adds to the width of a continent or island arc and stores material for longer periods. Large accretionary prisms involve a margin configuration and plate convergence constant for tens of millions of years. They are common where sediment fill in the trench axis is more than 1 km (0.6 mi) thick and rates of orthogonal convergence are less than 5 km/m.y. Examples are found along the Lesser Antilles and the Nankai, Makran, and Indonesian margins.

The core framework is commonly referred to as basement, and it includes diverse igneous as well as metamorphic and occasional sedimentary rock bodies. The framework basement exposed along coasts is a strong and rigid rock mass that can be followed tens of kilometers seaward with geophysical methods and sampling. Along more than half of the global length [about 41,000 km (25,000 mi)] of convergent margins, the framework extends to within 10–15 km (6–9 mi) of the trench axis. Many of the sedimentary rock bodies of the framework basement are fossil accreted wedges, as along the Japan and Alaskan margins.

The slope apron is a sediment sequence that overlies the margin framework and accreted prism. It extends from the edge of the shelf to the frontal prism and can be from a few meters to 5 km (3 mi) thick. The steepness of the continental slope causes gravity sliding and a general downslope migration of sediment. Seismic imaging of slope-apron stratification and high-resolution sea-floor bathymetry [100 m (330 ft) or less], show the lack of active deformation that indicates the tectonic stability of the margin framework and accreted prism relative to the frontal prism. Stability is also indicated by a sea-floor morphology with integrated canyons that are disrupted and disappear over the frontal

prism.

The plate interface is underlain by a layer from about 200–1200 m (about 660–4000 ft) in thickness called the subduction channel. This layer consists of subducted trench sediment and eroded clastic material on the igneous ocean crust and is commonly detected as a zone of low seismic velocity, relative to its bounding rock masses. The low velocity is interpreted as indicating a fluid-rich unit. Subduction beneath the frontal prism inflicts a load on the porous sediment in the subduction channel, which elevates the pore fluid pressure because the prism's permeability does not allow the fluid to escape as quickly as the load is applied. Fluid pressure can approach the weight of the overlying material. Such pressures greatly reduce friction and allow less consolidated materials to be subducted rather than scraped off at the deformation front. Farther down, the subduction zone temperature increases to levels where some clay minerals alter and release chemically bound water into the subduction channel. Elevated pressure in a fluid-rich layer is an explanation for reduced seismicity, and the manner of slip along the plate interface thrust is considered stable. The unstable slip released during large earthquakes and tsunamis occurs at ~10 km (6 mi) or greater depths in the portion of the subduction zone called the seismogenic zone. See also: Seismology; Tsunami

The subducting lower plate commonly bends into the trench axis and breaks into extensional or normal faults. This faulting produces voids that fill with sediment carried down the subduction zone on the lower plate. Some normal faults are imaged seismically through the entire oceanic crust, and could form aquifers that conduct seawater into the lower crust. Seawater in contact with lower crustal rock can change the mineralogy to weaker serpentine minerals and affect the mechanical behavior and geochemical evolution of the subducting lithosphere. The cold, negatively buoyant subducting plate becomes a driving force for convection in the Earth's mantle, which is the layer between the lithosphere and the core. Ultimately, subduction of oceanic plates into the mantle is responsible for much of the tectonism expressed at the Earth's surface.

Tectonic processes

The dynamic character of convergent margins is evident from earthquakes and tsunamis. Convergent margins begin in the seismogenic zone at depths beyond the reach of direct observation even with scientific drilling. However, the mechanisms involved leave a geology to interpret that indicates the nature of dynamic processes involved. Accretionary mass addition has an erosional mass subtraction counterpart. The geology of material added to the upper plate is more easily observed than the record of massive material removed from a continent or island arc and carried out of sight. An important but difficult record to decipher from geologic evidence is the role of fluid in controlling tectonism and seismicity.

Erosional convergent margins

Subduction erosion is predominant along more than 60% of convergent margins globally. At erosional convergent margins, the net volume of the margin's upper plate decreases during subduction. Evidence of missing rock masses includes finding (1) the roots of ancient volcanic arcs along the coast or more than 100 km (60 mi) closer to the trench axis than where they formed; (2) large-scale and long-term subsidence to trench depths of unconformities that formed in shallow-water environments; and (3) the migration of volcanic arcs hundreds of kilometers inland, indicating that the continental slope retreated a similar distance. Drill cores at seven Pacific convergent margins show 3–5 km (2–3 mi) of subsidence during the past 15–20 m.y. This subsidence requires upper-plate thinning from erosion along the plate interface because sediment deposition during subsidence rules out regional sea-floor erosion.

Erosional margins characteristically occur where plate convergence is relatively rapid (more than 4 km/m.y.) and sediment in the trench axis is less than 1 km (0.6 mi) thick. The lower plate of igneous oceanic crust and sediment bends downward into the trench axis, causing extensional structure diagrammed as tilted blocks or half-grabens (depressed, faulted blocks). The wedge-shaped upper plate consists of a continental crustal framework covered by a slope sediment apron. The margin framework is as strong at the coast, but downslope it fractures and develops extensional faults that break up the plate beneath the middle slope. Disruption of the framework structure results in mass wasting of the slope apron and basement. Debris from mass wasting migrates downslope and collects in a frontal prism beneath the lower slope. This thin weak apex of the frontal prism displays ridges and valleys at the sea floor that mimic the buried lower plate relief. A thrust fault bounds the underside of the frontal prism, but landward the plate interface becomes a subduction channel. In seismic images, the channel has a low-frequency-high-amplitude seismic reflectance that is 0.5–1.0 km (0.3–0.6 mi) wide, inferred to indicate a fluid-rich zone. Subduction beneath the frontal prism elevates fluid pressure in the lower plate and its sediment cover, and this pressure increases upward fluid migration along and across the plate interface. As fluid drains from the lower plate, its porosity is reduced, which consolidates and strengthens material below the interplate thrust, and at the same time fluid invading fractures in the upper plate weakens rock above the plate interface. Thus, the zone of minimum strength along the plate interface shifts upward. The shift results in a transfer of upper-plate material to the subducting lower plate, thereby thinning the plate. See also: Graben

In the subduction erosional model, the upper-plate framework loses strength downslope as the plate thins and fractures dilate from invasion by pressured fluid. As the lower plate subducts, it continues bending, which increases lower-plate relief. Relief promotes fracturing in the upper plate from jostling as lower-plate relief subducts beneath it. As hundreds of kilometers of lower plate subducts, the jostling and dilation of fractures by pressurized fluid ultimately fragments the base of the upper plate. Detached fragments are dragged piece by piece onto the underthrusting lower plate, which carries them into the seismogenic zone.

Accretionary margins

Accretionary margins preferentially occur where convergence is less than 40 km/m.y. (25 mi/m.y.) and where sediment in the trench axis is thicker than 1 km (0.6 mi). They are distinguished from erosional margins by the accreted prism that develops from the frontal prism. The sediment wedge filling the trench axis is deposited rapidly and has a high initial porosity (about 70%). As the wedge converges against the upper plate, horizontal force creates small compressional faults, which do not extend to the sea floor, in the lower part of the trench-sediment wedge, and pore fluid pressure increases. Before faults break the sea floor, fluid drains from the trench-sediment wedge through intergranular permeability. Closer to the deformation, front faults reaching the sea floor provide more efficient dewatering channels. The deformation front develops when upper sediment layers consolidate sufficiently to break in thrust faults. Faults near the deformation front may extend through most of the wedge sediment section, but a horizontal fault above the igneous basement and parallel to bedding (a decollement) that forms the interplate thrust fault intercepts them (**Fig. 11**). Sediment below the thrust fault subducts beneath the upper plate as part of the lower plate. Where it has been drilled, subducted sediment is little deformed and indicates low interplate friction, along with other observations. Once faulting landward of the deformation front begins, fluid drains rapidly through fracture permeability in the faults. The majority of original pore fluid in accreted sediment (80%) is expelled in approximately the first 5–10 km (3–6 mi) of the accretionary frontal prism. Accretionary frontal prisms thicken rapidly by thrust faulting, and it is here that the greatest permanent strain deforms sedimentary bodies. In the subducted sediment and igneous crust, fluid is released farther down the subduction zone.

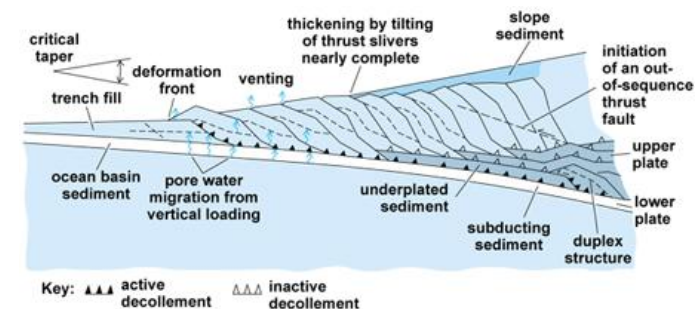


Fig. 11 Section across an accretionary convergent margin. The deformation front marks the partitioning of subducted and accreted sediment, a division that commonly occurs along the base of the trench fill. Rapid loading squeezes water from voids in the sediment, ultimately draining with the water from vents at the sea floor.

In many accretionary margins, the latest seismic and high-resolution bathymetric data show segmentation of tectonic activity not previously noted. A change in tectonic character separates the accretionary frontal prism from the accreted prism. The rapidly deforming frontal prism has a rough sea-floor morphology. The accreted prism deforms minimally, allowing deposition of a little deformed slope apron across it. The accreted prism is stronger and more rigid than the frontal prism, and the contact between them is a fault zone that commonly has sea-floor expression. This segmentation is not recognized along all accretionary margins, and it is difficult to resolve without the latest high-resolution geophysical data. Along the Nankai margin off Japan, the boundary between frontal prism and accreted prism is termed a splay fault, and similar features are resolved along the Sunda and Makran margins.

The concept of critical taper helps explain the thickening of an accretionary wedge. Taper refers to the angle between the upper and lower prism surfaces, and it depends on the strength of the material and the interplate friction. When the critical taper is exceeded, the slope becomes overcritical and fails by slumping to restore the optimum taper. When taper is undercritical, accretion restores it to the optimum taper for conditions at the time. To accomplish tectonic thickening, the accreted thrust packets are stacked and rotated, but only to a limiting angle. Once rotation and stacking of the detached packets can no longer thicken the wedge, thickening can be accomplished by the addition of material from below (underplating), by subsequent thrust faulting at a lower angle across previous structure (out-of-sequence thrusting), or by thrusting that is contained within a lower layer (duplexing). These processes occur at depths where clear seismic imaging of structures is at the limits of the current technology. They are currently modeled to explain taper related to upper plate thickening landward of the frontal prism.

The mass flux at convergent margins is important for understanding arc volcanism and the origin of continental crust. Subduction zones return crust to the mantle from which is once originated, but in an altered state. The past tendency to view subduction zones as primarily areas of growth has diminished as the extent of subduction zone erosion becomes known. The volume of continental crust returned to the mantle is small compared to its entire volume, but sufficient to impart the distinct continental geochemistry observed in mantle rock. Erosion is one argument for the paucity of very old exposed continental crust, functioning as a balancing process to have kept the growth of continents at their approximate size during much of the Earth's history (since pre-Cambrian time, about 2.5 billion years ago).

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