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THE OCEANIC LITHOSPHERE: SEISMOLOGY AND TECTONICS

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LECTURE 1 - PLATE TECTONICS AND SEISMOLOGY

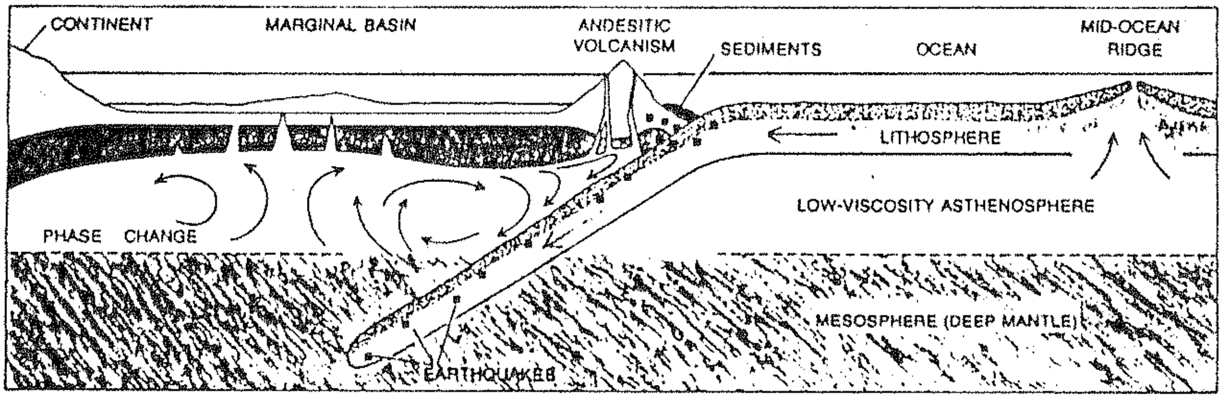
1. A quick overview of Plate Tectonics

The theory of Plate Tectonics presents the continents as being to a very large extent passive elements of global tectonics, while the ocean floor is the object of a much more rapid cycle; in this respect, the oceans are on the average much younger than the continents (and in no case older than 200 Ma), their geological history is far simpler, and thus their study gives easier insight into the fundamental tectonic processes involved both in the vicinity of the planet's surface and at greater depths. Thus, it should come as no surprise that Plate Tectonics was developed only in the late 1960's, on the basis of a wealth of oceanographic data which had been unavailable to previous generations of Earth Scientists.

The fundamental pattern of our present view of the evolution of the oceans, sketched on Figure 1, can be summarized as follows: New oceanic lithosphere is constantly being generated at the *Mid-Oceanic Ridges* [MOR], whose "spinal chord" extends in all three oceans. This continuous mountain range, by far the longest and largest tectonic feature on the surface of the Earth, is the locus of upper mantle upwelling, resulting, on either side of the MOR, in the accretion of basalts into rigid oceanic "plates", at the rate of a few (1 to 10) cm/yr. Upon formation, the oceanic plates are quickly cooled and move about the surface of the Earth obeying Euler's kinematics of rigid caps on a sphere.

At the same time, along lines known as *Subduction Zones*, oceanic plates are absorbed into the mantle; this results in a perfect balance between new material created at the MOR's and old oceanic floor recycled at the subduction zones, and keeps the radius of the Earth a constant. Subduction zones appear in the bathymetry as trenches, reaching as deep as 11 km; with the exception of Antarctica and a few other coastlines, subduction zones circumvent the Pacific, and are also found along the Indonesian arc; the Atlantic is practically deprived of subduction zones. As oceanic plates sink into the mantle, they remain detectable as slabs characterized by different physical properties (generally resulting from their colder temperature) as far deep as 670 km. This subduction is accompanied by a number of chemical reactions, due to the different chemistry of the slab and its saturation with water; they result in partial melting near the slab-mantle interface and the subsequent ascension of blobs of melt, inducing volcanism on the far side of the trenches, and thus the creation of arcs of volcanic islands. Active volcanoes are indeed known along all subduction systems.

While the oceans are involved in this large scale cycle of accretion at the ridges, cooling away from them, and eventual recycling through subduction into the mantle, the continents play the passive role of rafts. This is made possible by their granitic crust being lighter than the oceans' basalts, thus preventing subduction. Since they do not subduct, continents preserve the geological record of the planet since their original formation; of course, this record is made extremely complex by a succession of geological processes (e.g., erosion, metamorphism, cratering by meteorites,...), yet rocks as old as 3.8×10^9 yr [3.8 Ga] have been found in Greenland. Finally, when subduction rates in



FORMATION AND SUBDUCTION OF LITHOSPHERE are shown in this cross section of the crust and mantle. New lithosphere is created at a mid-ocean ridge. A trench forms where the lithospheric slab descends into the mantle. Earthquakes (*small* squares) occur predominantly in the upper portion of the descending slab. Arrows in soft asthenosphere indicate direction of possible convective motions. Secondary convection currents in asthenosphere may form small spreading centers under marginal basins.

Fig. 1*a*: Sketch of the evolution of the oceanic lithosphere, from its formation at the Mid-Oceanic Ridge to its subduction at the trench [*Toksöz*, 1976].

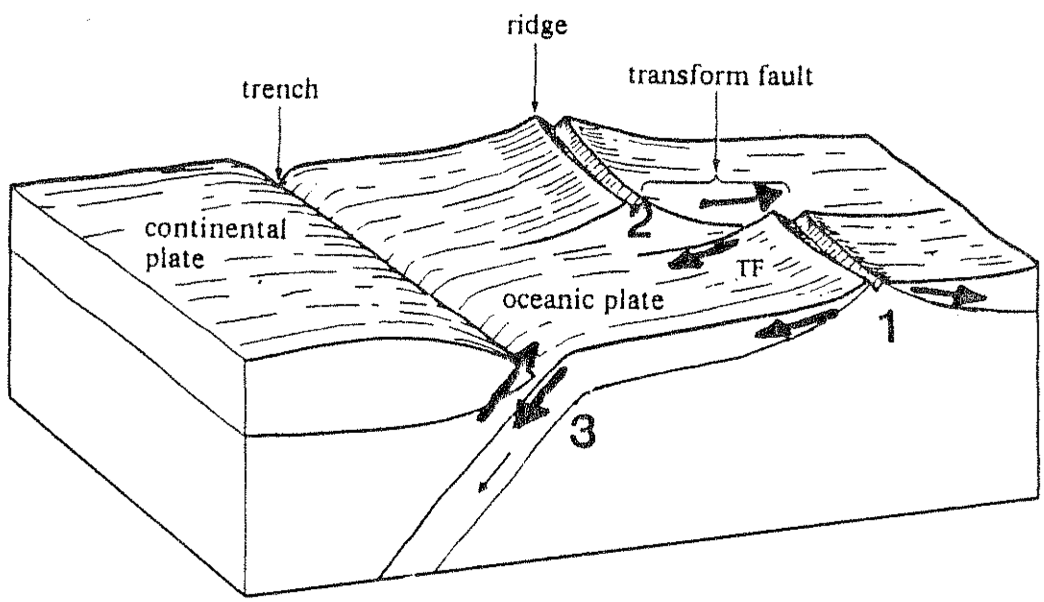


Fig. 1*b*: Simplified kinematics of an oceanic plate. Arrows and numbers refer to the three types of earthquakes at plate boundaries described in text. After *Forsyth and Uyeda* [1975].

a given ocean are faster than accretion, the whole ocean is doomed and eventually continental rafts collide. Since they cannot subduct, gigantic denting takes place in the form of mountain building. This process is, however, a slow one: for example, it has taken 38 Ma of India ramming into Southern Asia to build the Himalayas, lift Tibet 5 km up, pop the Tien Shan Range up the Asian continent and start squeezing China out into the Pacific Ocean. As such, the most spectacular vertical tectonics observed on continents are clearly by-products of a much larger cycle of horizontal processes, operating one order of magnitude faster (a few cm/yr rather than a few mm/yr), but camouflaged by an average 5 km of oceanic water. Occasionally, entire sections (e.g., slivers) of the oceanic crust can be incorporated into mountain ranges during the collision process. These formations of ancient oceanic rocks presently exposed on continental areas are known as *ophiolites*, and are extremely valuable to our understanding of the oceanic lithosphere. Spectacular ophiolites exist in Oman, Cyprus, Newfoundland and California.

2. The evidence which led to Plate Tectonics

On the basis of data borrowed from coastal geography, structural geology, paleontology and paleoclimatology (all of which came from continental observations), *Wegener* [1915] had boldly proposed the drifting of continents. In the absence of oceanic evidence, his model met with harsh criticism, and by the time of his death in 1929, his ideas were to a large extent repudiated by a somewhat biased scientific community.

The main contributions to putting together the puzzle of accretion at the MOR's and eventual subduction, came from straight bathymetry, paleomagnetism and radiometric dating. In particular, one of the crucial studies was that of the pattern of magnetic anomalies in the world's ocean basins. These small-scale fluctuations of the intensity of the Earth's magnetic field were found to be grossly symmetric about the axis of the MOR's, and correlated not only from ocean to ocean, but also with magnetostratigraphic columns worked out from continental rocks. The giant leap forward came with *Vine and Matthews'* [1963] realization that the anomalies were the magnetic record of the Earth field's reversals, and that their geometrical layout was linked to a progression of the age of the oceanic floor away from the MOR's, and thus stemmed from *seafloor spreading*.

3. Seismicity fits the puzzle

Seismology is the study of earthquakes and of the waves they generate. An earthquake is a sudden readjustment of Earth masses, causing relative motion of blocks of rock with characteristic times at most comparable to the Earth's elastic eigenperiods (less than an hour, and in practice much shorter). Seismic waves are just about the only form of propagation through the Earth's interior which is not severely damped after a short distance; thus the study of the propagation of seismic waves has been the fundamental tool of our investigation of the properties of the Earth's interior. The next lecture will describe these techniques and their application to the particular case of the oceanic lithosphere. On the other hand, the branch of Seismology investigating the occurrence of earthquakes has brought considerable insight into the nature of the internal dynamic processes of the Earth.

After reliable electromagnetic seismographs were introduced in the early 1900's, and following several decades of careful compilation, a clear picture of the Earth's seismicity was available by 1940. First, all earthquakes occur in the shallowest portions of the Earth. No seismic event has been detected below 700 km depth, and careful relocation of the deepest events indicates maximum depths of 670 km. In addition, the repartition at the surface of the Earth of the radial projection of earthquake sources (*epicenters*), is clearly arranged along lines of preferential seismicity (see Figure 2). We now recognize them as being the *plate boundaries* along which subsequent earthquakes

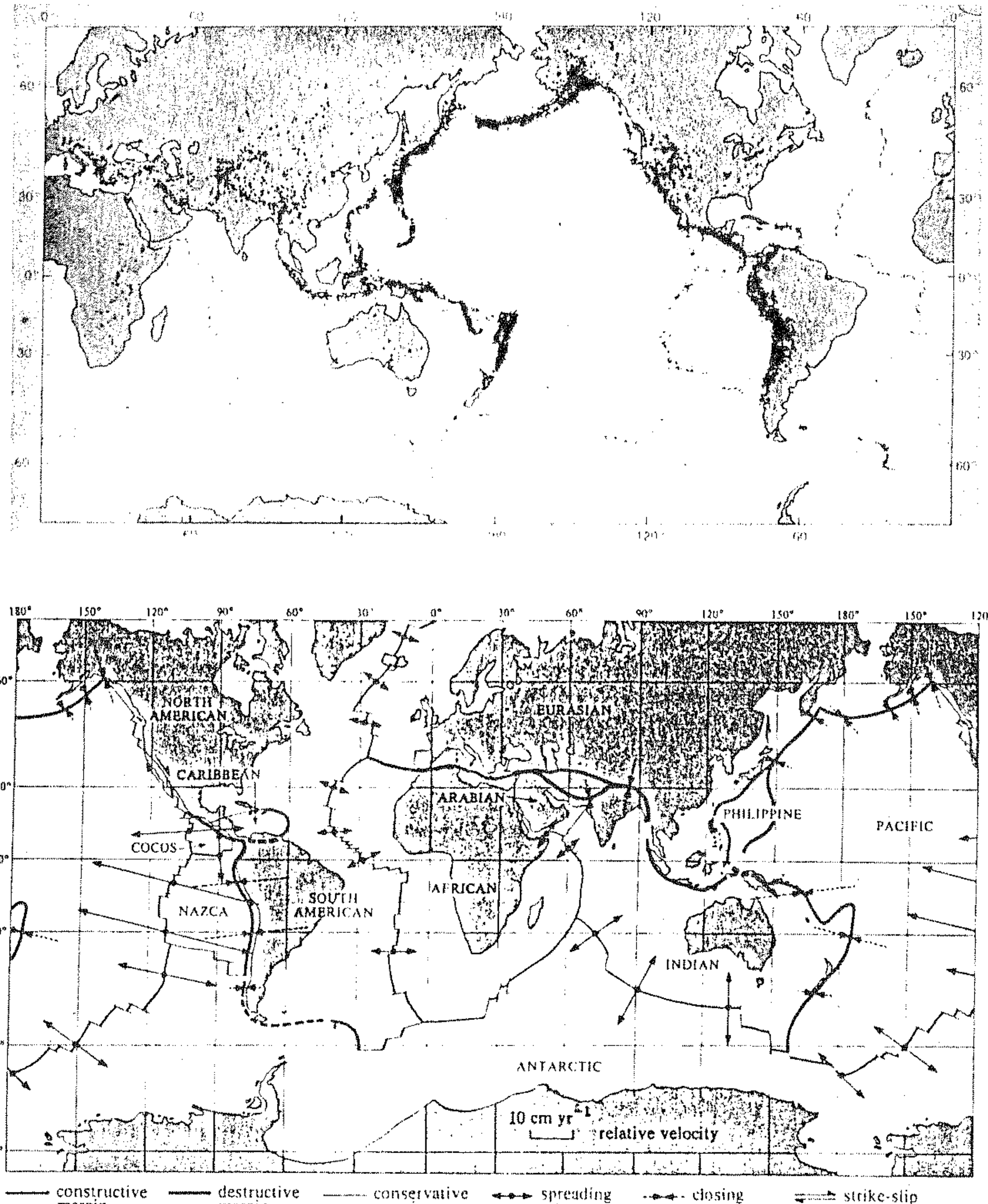


Fig. 2: Top, world seismicity map. Note concentration at Mid-Oceanic Ridges and Subduction zones. Bottom, Geometry of the principal plate boundaries, with nature and amplitude of relative velocity vectors [*Brown and Mussett*, 1981].

contribute to the relative motion of a system of two plates.

Seismological techniques based upon obtaining a worldwide dataset of records from a moderate-to-large earthquake, allow the routine reconstruction of the geometry of the faulting blocks at its source. Their systematic application to the plate boundary system by *Isacks et al.* [1968] has confirmed the existence of three basic kinds of plate boundaries (see Figure 1*b*):

(1) Along MOR's, the constant extrusion of volcanic material to accrete additional plate elements, is accompanied by tectonic processes such as the formation of valleys and grabens, which may involve normal faulting earthquakes, expressing the extensional nature of the process. This is the case of such slow-spreading ridge systems as the Mid-Atlantic Ridge (2 cm/yr), or the Southwest Indian Ocean Ridge. Note however that fast-spreading systems, such as the East Pacific Rise (18 cm/yr) are deprived of ridge crest earthquakes (at least at the worldwide level of teleseismic detection of $m_b = 4.7$; we presently do not know if smaller earthquakes exist on fast ridges).

(2) The MOR's are occasionally interrupted by *Transform Faults*, laterally offsetting the spreading segments by as much as 1000 km. Along them, earthquakes feature a strike-slip mechanism, simply expressing relative motion of the two plates parallel to their boundary. The understanding of the nature of a transform fault by *Wilson* [1965] was a key step in the development of the theory of sea-floor spreading. It is remarkable that the only seismological evidence available to Wilson at the time, was the absence of earthquakes outside the transform segment of a FZ. Seismic studies such as Isacks *et al.*'s later provided a spectacular confirmation of Wilson's model, from focal mechanisms of transform fault earthquakes (the transcurrent fault model would predict exactly the opposite motion). In California, the San Andreas fault is interpreted as a transform segment of the East Pacific Rise system, quickly approaching total subduction under North America, but still recognized off the coast of Oregon and in the Gulf of California.

(3) At most subduction zones, seismicity involves very large earthquakes characterized by a thrust faulting mechanism, expressing the overriding of the subducting oceanic plate by the other plate. Most of the large destructive earthquakes around the Pacific are of this type (e.g., Chile, 1960; Alaska, 1964; Aleutian, 1965; Japan, 1968).

All three types of seismic events described above are characterized by extremely shallow depths. Ridge-crest or Transform fault events are usually less than 15 km deep; subduction events occur mostly above 70 km. We will see in the next lecture that the typical thickness of the plates, as revealed by seismic waves, is 100 km; thus only the shallowest sections of the plates are seismically active: at greater depths, the temperature inside the plate rises quickly, rocks are no longer brittle and tectonic processes are expressed through ductile fracture, which does not generate seismic waves.

Under subduction zones, however, and because the subducting slab remains *colder* than the adjoining material, deep earthquakes can and do exist, down to 670 km. These events do not express motion between two plates, but rather the release of stresses which build up inside the slab, as it penetrates the mantle. They are confined to the slab, since it is the only environment in which brittle rupture can take place at these depths. Thus, plots of seismicity as a function of depth across subduction zones allow a spectacular *mapping* of the descending slab; Wadati and Benioff used them to gain the first evidence of intrusion of lithosphere into the mantle, practically the only contribution of Seismology to the early development of the concepts of Plate Tectonics.

Finally, elaborate analytical techniques allow the retrieval from seismological data of the actual *amplitude* of the rupture displacement at the earthquake source. If applied to areas of the world with sufficient historical records, they can be used to obtain an average rate of seismic displacement, which is then compared to available models of plate kinematics (themselves obtained from the study of young magnetic anomalies in the immediate vicinity of the MOR's). In general, these results have indicated that earthquakes account for a significant, and occasionally major, part of the motion between plates. But on the other hand, an important fraction of the plates' motions escape seismic detection, and must take place through processes such as extremely slow motion, or permanent creep. More specifically, spreading and accretion at MOR's is only marginally seismic (totally aseismic at fast MOR's); along most oceanic transform faults, earthquakes give a near-perfect account of the plates' motion, although some exceptions exist (e.g., the Eltanin Fracture Zone in the South Pacific). Subduction zones can behave as a strongly coupled system totally released through seismic events (e.g., Chile), or be largely decoupled, with a paucity of large earthquakes (e.g., the Mariana Trench area) [*Uyeda and Kanamori*, 1979]. The situation along the San Andreas Fault is more complex, with the Northern and Southern sections seismically active, and the central one decoupled and creeping.

LECTURES 2 & 3 - SEISMOLOGY AND THE STRUCTURE OF THE OCEANIC PLATES

1. Seismic waves

The elastic waves radiated from a seismic source into an elastic medium, such as the Earth, consist of *body waves*, propagating through the interior of the Earth, and *surface waves*, creeping along the planet's surface. Figure 3 shows an example of the succession of body and surface waves generated by a large earthquake. Body waves are probably the easiest to visualize, since they obey Fermat's principle of stationary path, and more generally speaking, geometrical optics. The situation is made somewhat complex, however, by the existence of 2 kinds of seismic body waves: P waves, corresponding to an elastic disturbance in the direction of motion, and involving a change of density in the medium, and S waves, corresponding to a vibration perpendicular to the direction of motion, and involving only a shear deformation of the medium, without compression or dilatation. The propagation of P waves is similar to the vibration modes of an elastic *spring*, while that of S waves resembles the vibration of a *string*. The velocities of propagation of P and S waves are, respectively:

$$\alpha = \left[\left(K + \frac{4}{3}\mu\right) / \rho \right]^{1/2}, \text{ and}$$

$$\beta = \left[\mu / \rho \right]^{1/2},$$

where ρ is the medium's density, K its incompressibility and μ its rigidity. Since K and μ cannot be negative (from thermodynamic arguments), α is always greater than β, i.e. P waves travel faster than S waves. In practice, and for most solids, the ratio α/β is between 1.5 and 2.0 .

In the case of liquids (or gases), the medium has no rigidity (a liquid cannot stand by itself and needs a container), μ is zero, and there exist no S waves. Then P waves are just the familiar sound waves.

Another interesting structural parameter is the degree of elastic imperfection of the medium. As a seismic wave propagates through a real medium having some anelasticity, a fraction of its energy is converted by friction into heat (mostly through grain boundary relaxation) and dissipated, resulting in a steady decay of the wave's amplitude with time (or distance). The attenuation of the wave is characterized by a quality factor Q defined by

$$\frac{1}{Q} = \frac{1}{2\pi} \Delta W / W,$$

where $\Delta W/W$ is the relative loss of seismic energy over one period of the wave.

The (relatively difficult) measurement of the attenuation of seismic waves allows the recovery of the structural anelasticity of the internal layers of the Earth. It has been found experimentally that 2-phase media, in particular materials containing small liquid inclusions (such as those resulting from the initiation of partial melting) have lower values of Q.

2. Body wave methods

As one progresses into the Earth's interior, and as a general rule of thumb, seismic velocities *increase* with depth. This is due partly to a change in the nature of the rocks, and also to the steady increase of pressure with depth, with elastic constants, such as K and μ increasing with pressure faster than ρ. Because seismic rays obey Fermat's principle, in a medium where velocity increases with depth, their path is curved downwards, and generally speaking they travel deeper than the straight line. It is thus possible to recover Earth velocities at depth with both source and receiver at the surface, e.g., on or near a ship in the marine environment.

The interpretation of travel time curves resulting from a seismic refraction experiment is simplest under the assumption of *lateral homogeneity*, i.e., that the structural properties of the medium depend only on one coordinate. This is usually sufficient for small-scale experiments. In the case of a flat-layered Earth modeled as a discrete number of homogeneous layers, the first arrival travel-time curve is made up of a number of straight segments (see Figure 4). The slopes of the segments are the inverse of the successive velocities, while the positions of their intercepts are related to the layers' thicknesses. In the more general case of a continuously varying, but still laterally homogeneous medium, the seismic velocity as a function of depth $c(z)$ (c can be either α or β) is recovered from the observed first arrival travel-time as a function of distance $T(x)$ by the following method, due to Wiechert and Herglotz:

- Define $p(x) = \mathrm{d}T(x)/\mathrm{d}x$
- For each value X of the distance, compute the following integral over all values $x \leq X$:

$$z(c) = \frac{1}{\pi} \int_0^X \cosh^{-1} [p(x)/p(X)]\, dx$$

- Then, z is the depth at which the seismic velocity takes the value $c = 1 / p(X)$.

The method can also be applied to the case of spherical symmetry, when the range of the experiment becomes comparable to the radius of the Earth a. In this case, one obtains experimentally a travel time function $T(\delta)$, where δ is the angular distance between source and receiver. The Wiechert-Herglotz method then goes through the following steps:

- Define $p(\delta) = \mathrm{d}T(\delta)/\mathrm{d}\delta$
- For each value Δ of the distance, compute the following integral over all values $\delta \leq \Delta$:

$$y = \frac{1}{\pi} \int_0^\Delta \cosh^{-1} [p(\delta)/p(\Delta)]\, d\delta$$

- Then, at the radius r (or the depth $a - r$), given by

$$r = a\, e^{-y}$$

the seismic velocity takes the value $c = r / p(\Delta)$.

A representation of the function $c(r)$ is thus obtained, parameterized by the variable p. Note however that in order for the function $\cosh^{-1}$ to be real, p has to be a constantly decreasing function of x. By the same token, note that the resulting function

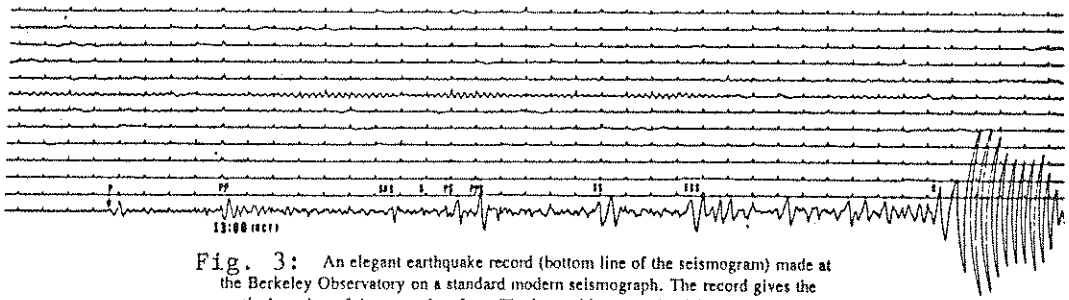
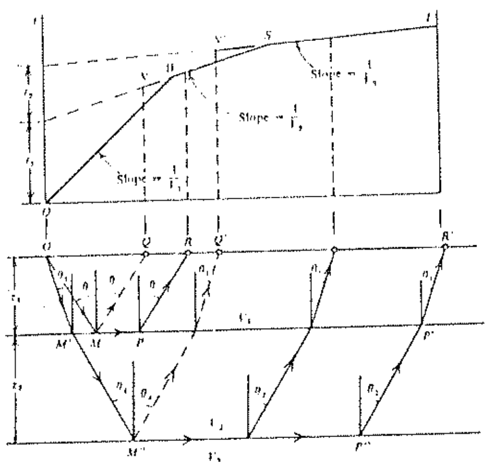


Fig. 3: An elegant earthquake record (bottom line of the seismogram) made at the Berkeley Observatory on a standard modern seismograph. The record gives the vertical motion of the ground surface. The interval between the tick marks on the record corresponds to 1 minute, and time increases from left to right. This earthquake occurred near Borneo at a distance of 11,000 km from Berkeley. The onset of *P* waves is clearly seen, together with the *PP* reflection. This is followed by the onset of *SKS* and *S* waves and the reflections *PS*, *PPS*, *SS*, and *SSS*. At the end of the bottom trace can be seen the Rayleigh wave train, starting with long but decreasing periods (an example of wave dispersion). The record is not complete because the wave motion was interrupted by the operator inadvertently changing the seismogram. [*Bolt*, 1982].



Raypaths and traveltime curves for two-refractor case.

Fig. 4: Principle of the interpretation of seismic refraction data, in the case of flat layering [*Telford et al.*, 1976].

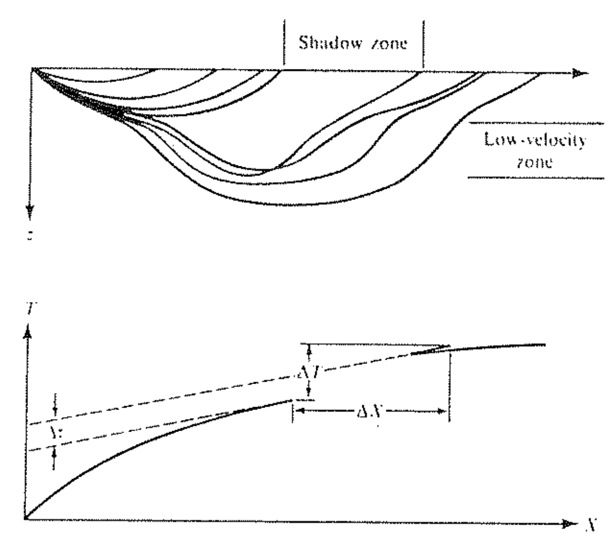


Fig. 5: Effect of a Low-velocity zone on the ray paths of seismic waves, and on their travel-times. Note the development of a shadow zone [*Aki and Richards*, 1980].

$z(c)$ is always increasing with increasing c, and thus that the function $c(z)$ cannot decrease with z. We anticipate that the method will not apply when a low-velocity layer is present in the structure.

In addition to this technique of *seismic refraction*, strong discontinuities in the Earth have different seismic impedances, and are capable of reflecting seismic energy. *Seismic reflection* is used extensively to recover shallow seismic structure, both for general and commercial exploration.

3. Results from body wave methods; shallow structures.

Seismic sources used in the investigation of the structure of the oceanic basins have included both natural sources (earthquakes) and man-made explosions of a widely variable yield.

The oceanic column, whose thickness varies from a typical 2000 m at the MOR's to 5.5 km in the older oceanic basins, has a sound velocity of 1.5 km/s. Slight variations around this value are controlled by temperature and salinity. Precise studies have shown the existence of a channel of low-velocity ($\alpha \sim 1485$ m/s, as opposed to 1520 m/s at the surface) at depths ranging from 600 to 1800 m. Since very little attenuation takes place in the oceanic water, waves trapped in this guide, known as the *SOFAR* channel can propagate over very long distances. This propagation is efficient only at wavelengths much shorter than the thickness of the *SOFAR* (typically at periods less than 0.5 s), and the corresponding waves, known as T waves, have extraordinary long-distance detection capabilities (they have been used both for tracking submarines and detection and monitoring of active underwater volcanoes).

Below the oceanic column, the following layers have been recognized in the early 1950's [*Raitt*, 1963].

TABLE 1. Typical Seismic Structure of the Oceanic Crust

Layer	Thickness, km	P-wave velocity, km/s
Ocean	4.5	1.5
1	0.45	2
2	1.75	5
3	4.7	6.71
4	-	8.09

4. Interpretation

Although in principle, it is possible to interpret seismic profiles by comparison with data obtained on laboratory samples, recent developments in oceanic drilling (e.g., the DSDP) and the study of ophiolites have helped us considerably in obtaining insight in the nature of the various layers evidenced from seismology.

Layer 1 is identified only from reflection. It is very thin, and features P wave velocities only slightly greater than that of water. This layer is made of unconsolidated sediments, as confirmed by results of the DSDP.

Layers 2, 3, and 4 are detected mostly from refraction arrivals. While the general layering is consistent in various areas of the oceans' basins, the thickness and velocities inside the individual layers exhibit wide variations : e.g., layer 2 has been found faster than average near some groups of seamounts and islands. It was argued for a while what the exact nature of Layer 2 could be - consolidated sedimentary rocks (limestones) or basaltic lavas. The final evidence came from the eventual drilling of Layer 2, and indicated that it is indeed made of basaltic rocks similar to those cropping at the MOR's where sediments have not had the time to accumulate. Layer 2 is believed to be the main contributor to the familiar magnetic field anomalies of the oceans' basins.

Layer 3 is the main oceanic crustal layer, extending for a typical thickness of 4-5 km, and with high seismic velocities reaching up to 7.6 km/s. The nature of Layer 3 was also argued for a while, between gabbroic and serpentinite compositions; serpentinite being a highly hydrated phase features anomalously low ratios of β to α; this is not observed in Layer 3, and this favors the gabbro hypothesis. Also, in view of the recent discovery of hydrothermal circulation near the ridges, recent models of the crustal temperatures are generally higher, to the point where formation of gabbro is possible.

The boundary between Layer 3 and Layer 4, a sharp discontinuity about 7 km deep and featuring a jump to α = 8.15 km/s , is interpreted as the oceanic Mohorovičić [Moho] discontinuity. Thus, Layer 4 represents the top of the mantle, and from the study of ophiolites, we know that it consists principally of peridotites, mainly dunite and hartzburgite.

Obvious differences between oceanic and continental structure involve the much thinner crust, and the chemical nature of the transition at the Moho.

5. Limitation of body-wave techniques. Existence of the low velocity zone

A very stringent limitation of the Wiechert-Herglotz method is the fact that it yields depth as a function of velocity rather than the opposite (this is most clearly seen in the flat-layered framework). The method then works well if velocity increases monotonically with depth, but breaks down if a *low velocity zone* is encountered. In such a case, a decrease of velocity with depth results in a steepening of seismic rays (according to Snell's law, a consequence of Fermat's principle), and the development of a *shadow zone*, i.e., a range of distances at the Earth's surface where no seismic geometry obeying geometrical optics is received (see Figure 5). It is clear that the Wiechert-Herglotz integral over x (or δ in the spherical case) cannot be carried out.

As early as 1926, B. Gutenberg hinted at the existence of such a shadow zone, for oceanic seismic paths between 6° and 16°, and had it firmly established by 1948 [*Gutenberg*, 1926; 1948]. On the other hand, travel time data along paths sampling stable continents (e.g., shields) do not feature a shadow zone. This was the first evidence for heterogeneity in deep structure between continents and oceans.

The breakdown of the Wiechert-Herglotz method at about 100 km depth in the oceans requires the use of a different approach.

6. Surface wave methods

Two types of seismic waves can propagate along the Earth's free surface: In *Rayleigh* waves, the seismic motion of a point at the surface of the Earth is a small ellipse whose major axis is vertical, and minor axis horizontal in the direction of propagation. This motion involves both shear and compression, and can be thought of as the superposition of a P wave and an S wave, both having imaginary components to their wavevector. In *Love* waves, the motion is horizontal, at right angles to the direction of propagation, and involves only shear; thus Love waves cannot penetrate a liquid, and "oceanic Love waves" involve only the ocean's floor. Love waves can exist only in layered media, or in media with curved symmetry. The seismic displacement's amplitude in both Rayleigh and Love waves decays away from the surface, the characteristic depth of penetration being on the order of $(\lambda/4)$, where λ is the horizontal wavelength. This penetration is therefore frequency-dependent; conversely, waves of different periods will sample the Earth's structure at different depths. In a heterogeneous medium, this leads to a *dispersion* of the wave, i.e., a phase velocity $c = \omega / k$ varying with ω , and a distinct group velocity $U = \mathrm{d}\omega/\mathrm{d}k$.

Systematic measurements of both phase and group velocities of Rayleigh and Love waves having traveled over oceanic paths can be inverted to resolve Earth structure over depths comparable to a fraction of the wavelength. The study of the decay of the amplitude of surface waves with distance traveled similarly yields information on the degree of anelasticity in the Earth as a function of depth. Basic steps are as follows:

Phase velocities are obtained either along a path linking an earthquake and a seismic station, or between two stations aligned with the epicenter on a common great circle. This last technique has the advantage of eliminating the phase characteristics of the seismic source; the phase velocity is then retrieved from the difference in phase of the Fourier transforms of the records at both stations. Group velocities are obtained after narrow-bandpass filtering, by computing the group arrival of the maximum of energy of the wavepacket, which is obtained from the seismic record itself, and its Hilbert Transform.

The interpretation of the dispersion characteristics is primarily based on refining a "starting" model of the Earth. Theoretical dispersion curves are computed for this initial model, and residuals of the velocities computed. The dispersion calculations also yield the partial derivatives of the observables (c, U) with respect to the structural parameters (e.g., by how much the phase velocity of 70-s Rayleigh waves is changed when the shear velocity is increased 0.1 km/s between 200 and 300 km depth). These "partials" can be used to linearize the inverse problem, and methods of generalized inversion, developed for a number of geophysical applications, are then applied to the retrieval of a "best" Earth structure fitting the observed dispersion data. In practice, the shear velocity partials are several times larger than their density or compressional velocity counterparts. In other words, Rayleigh wave dispersion is mostly sensitive to the values of shear wave velocities in the crust and mantle. This is due to the fact that among the (complex) P and S waves making up the Rayleigh wave, the P wave decays about twice as fast with depth as does the S wave. Love waves are of course sensitive only to the shear velocity structure, and very marginally to density.

Finally, attenuation data can be inverted into intrinsic Q values in the Earth along exactly similar lines.

7. Systematic oceanic investigations

The earliest attempts at defining the structure of the oceanic upper mantle from surface wave dispersion data go back to the early 1960's. Following intense work by a number of scientists during 1963-1971, it was generally found that oceanic models require a strong low-velocity zone [LVZ], where the shear-wave velocity is approximately 0.5 km/s slow, and extending in gross numbers between 100 and 250 km. The LVZ was also identified as the primary zone of attenuation in the mantle.

It was recognized early on that continental structure is intrinsically heterogeneous (with for example no LVZ detected under old shields), and by 1970, regionalizations of most surface wave dispersions carried several continental regions [e.g., *Kanamori*, 1970]. It is not until 1974 however that the question of the lateral variation in structure of oceanic areas was addressed. *Leeds et al.* [1974] used paths crisscrossing the Pacific Ocean to study the variation of dispersion characteristics of Rayleigh waves. As shown on Figure 6*a*, it is clear that Rayleigh phase velocities are maximum in older areas of the oceans, and slowest along the paths remaining close to the East Pacific Rise, where the plate is very young. *Leeds et al.*'s [1974] inversions were carried out by letting only one parameter vary, namely the depth to the top of the LVZ. The general evolution toward faster velocities in old oceans resulted in a deepening of this parameter from 0 at or near the ridge to about 150 km in the oldest parts of the Pacific.

A few years later, *Mitchell and Yu* [1980], working from a considerably larger dataset, including Love as well as Rayleigh data, taking into account attenuation as well as dispersion, and allowing full variation of the shear-wave velocity profile, produced a series of models for the evolution of the structure of the oceanic upper mantle (see

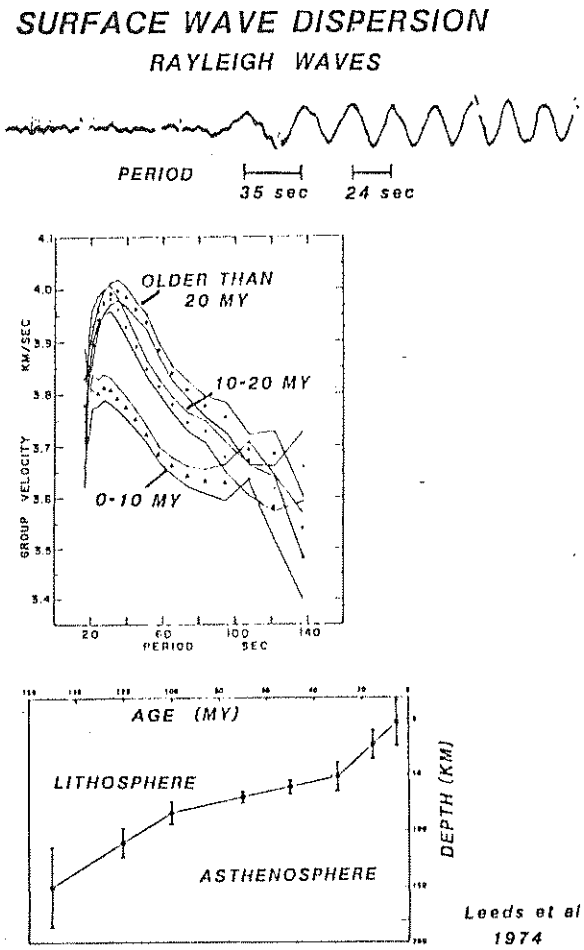


Fig. 6*a*: Influence of plate age on the propagation of surface waves. The top trace shows the dispersion of group velocity with frequency with 35-s energy preceding 20-s energy by about one minute. The center shows the variety of phase velocities obtained for different plate ages. The bottom diagram shows the result of *Leeds et al.*'s [1974] one-parameter inversion, featuring a thickening of the plate at the expense of the low-velocity channel.

Fig. 6*b*: Results of *Mitchell and Yu*'s more complete inversion, showing both the thickening of the plate, and the shallowing of the bottom of the low-velocity zone with age. In this figure, the solid line refers to the youngest model, followed, in order of increasing age, by the dashed, dash-dot, and dotted lines.

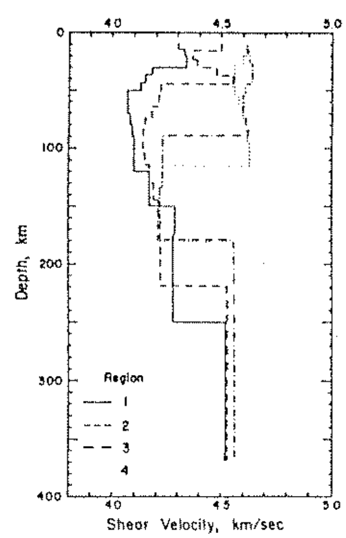


Figure 6*b*). Their basic results are as follows: (i) the depth to the top of the LVZ increases with age from about 0 to 120 km (ii) the depth to the bottom of the LVZ decreases from 280 to 170 km. (ii) the velocity in the LVZ remains on the order of 4.05 km/s. (iv) attenuation decreases with the age of the ocean, with the oldest regions having an average Q about 1.5 times that at the ridges.

These models obtained from regionalization of waves having traveled many portions of the Pacific, have been upheld by localized measurements of the Rayleigh dispersion in Polynesia by *Okal and Talandier* [1980]. The very deep extent of the LVZ in very young oceanic structures has also been confirmed in recent studies by *Montagner and Jobert* [1982], using very-long period Rayleigh waves in the East Pacific Rise area.

However, and very recently, *Anderson and Regan* [1983] have proposed that the LVZ may be as shallow as an average of 50 km under the oceans, on the basis of the effect of including intrinsic anisotropy in simultaneous inversions of both Love and Rayleigh data. If these models are confirmed by independent approaches, they could change significantly our interpretation of the dynamics of plate motion.

8. The plate concept

The most important element in the structure of the upper mantle under oceans is undoubtedly the LVZ. Its low velocity and high attenuation make it a medium with poor mechanical properties. It was recognized around the late 1960's that it provides the decoupling layer allowing motion of the oceanic floor above, and independently of, the rest of the mantle. As a result, the motion involved in sea-floor spreading and continental drift takes place only over very shallow depths, and the part of the Earth involved in the tectonic cycle of the ocean floor is very thin, and the term "Plate" Tectonics was coined around 1967.

The nature of the physical phenomenon leading to the existence of the LVZ is clearly related to the increase in temperature with depth in the mantle. However, temperature alone cannot explain the decrease in seismic velocities, the considerable increase in seismic attenuation, and the very limited range of depths where these properties are evidenced. The most widely accepted suggestion is that partial melting is present in the LVZ; laboratory experiments indicate that pressure and temperature conditions would be adequate for the development of partial melting below depths of ~100 km, in the simple case of an olivine mantle with traces of water (~0.1 %).

In summary, the oceanic plate, as determined by the study of seismic velocities is approximately 80 km thick (0 at the ridges, 120 km in very old ocean). It is cold, has high velocities, low attenuation (high Q), and also high viscosity. It rides on a LVZ, whose thickness varies from 300 km at the ridges to 70 km in the very old oceans, and where seismic shear velocities drop by 15 %. Most of the Earth's seismic attenuation takes place in the LVZ. Its viscosity is probably 2 orders of magnitude lower than that of the plate.[1]

9. Anisotropy: hints on the formation and evolution of the plate

Seismic refraction experiments in the mid-plate oceanic environment have shown that the velocity of propagation of seismic waves just below the Moho discontinuity (P_n) features *azimuthal* anisotropy, i.e. depends slightly on the azimuth of propagation; the amount of anisotropy was found to be approximately 4 %, and the direction of fastest velocity to coincide with the azimuth of fracture zones [*Raitt et al.*, 1969].

Comparison with a number of laboratory experiments has suggested that upper mantle seismic anisotropy most likely originates from a preferred orientation of

[1]Seismologists use the words lithosphere and asthenosphere for the plate and the LVZ, respectively. When dealing with phenomena featuring time constants much longer than the periods of seismic waves however, the term asthenosphere refers to a much shallower layer.

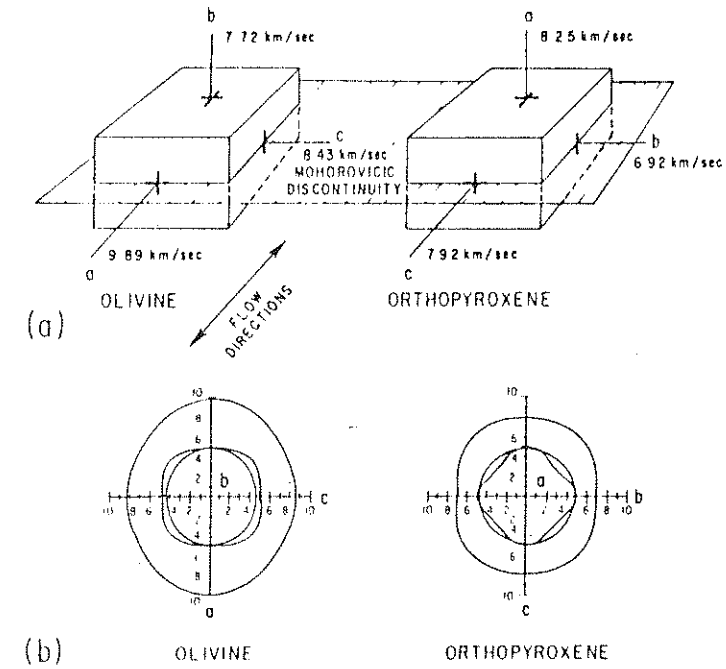


Figure 4. (a) Olivine and orthopyroxene orientations within the upper mantle. Single-crystal compressional wave velocities are for propagation along the crystallographic axes. (b) Compressional and shear velocities within the crystallographic olivine a–c plane and orthopyroxene b–c plane.

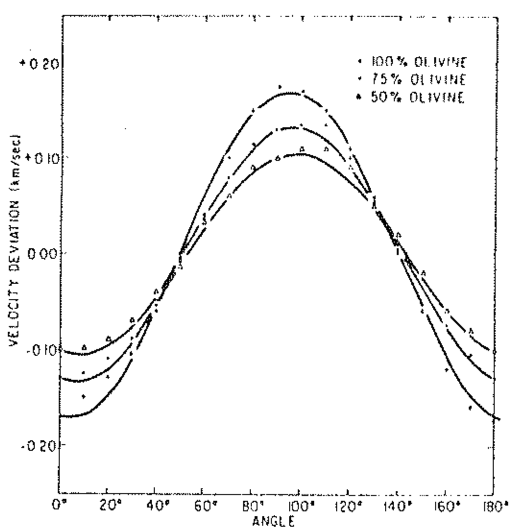


Figure 7. Anisotropy versus azimuth within the plane of the Mohorovičić discontinuity for varying proportions of olivine and orthopyroxene, using the calculated velocities shown in 5a. Inferred spreading direction is at 90°.

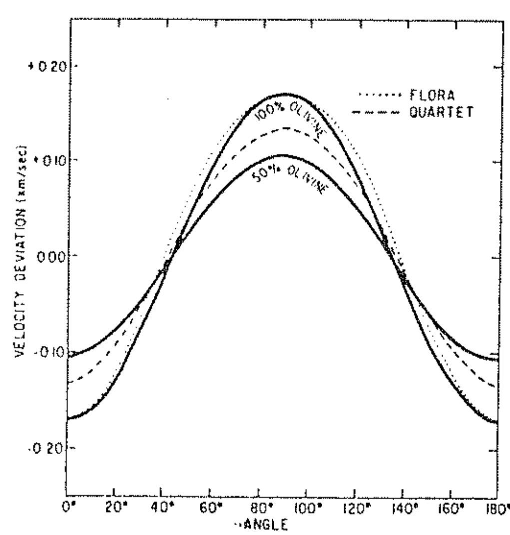


Figure 8. Comparison of upper-mantle seismic anisotropy from the Quartet and Flora areas (Raitt and others, 1967, 1969), with anisotropy calculated from petrofabric analyses of the ultramafic section in the Bay of Islands ophiolite.

Fig. 7: Anisotropy of the upper mantle strcture as inferred from petrography of ophiolite sections (top). The bottom diagrams present the expected seismic anisotropy computed for a variety of mixtures of olivine and orthopyroxene (left), and matches it to observations in the Pacific [*Christensen and Lundquist*, 1982].

crystals of olivine. The existence of preferred orientation is itself explained as a result of crystallization in the presence of stresses at the ridge, with the geometry of the crystals *frozen* into the plate at the time of its formation. The study of ophiolites, and in particular of sheeted dykes, have confirmed the preferred orientation of the fast axes of olivine crystals parallel to the inferred spreading direction.

More advanced models take into account the importance (up to 40%) of orthopyroxene as a component of the mantle layers of ophiolitic complexes [*Christensen and Lundquist*, 1982]. Detailed investigations of the crystalline orientation of olivine and pyroxene in these sections has revealed that the pyroxenes are emplaced with their fast axis vertical, and their slow axis parallel to spreading (just the opposite of olivine). As a result, the total amount of horizontal azimuthal anisotropy is a function of the volume percentages of the two components, and theoretical calculations can be used to retrieve the composition of the upper mantle (see Figure 7).

Azimuthal anisotropy has also been reported in the propagation of surface waves across oceanic plates. While the amount of anisotropy is similar, the direction of maximum velocity has been found to be that of present-day spreading (and therefore of present-day stress), rather than the direction frozen in at the time of formation. This apparent contradiction is explained by the fact that Rayleigh waves sample deeper parts of the mantle than do P_n. This deeper material may have a different preferred orientation, due either to its higher temperature, allowing its crystals to relax along the present-day stress, or to the fact that the deep layers sampled by Rayleigh waves may have crystallized more recently, at the bottom of the plate, and thus under the present-day stress conditions [*Okal and Talandier*, 1980].

LECTURE 4 - INTRAPLATE EARTHQUAKES AND THE STATE OF STRESS OF THE OCEANIC LITHOSPHERE

1. Introduction; intraplate earthquakes

In view of the success of the relatively simple ideas of plate tectonics, and especially of the interpretation of the major earthquakes as related to the relative motions between rigid plates, the mere concept of *"intraplate seismicity"* (see Figure 2) may sound like a paradox, if not like a failure of the whole theory. Fortunately, the relatively low level of this seismicity saves the concept of rigid plates, and allows its study as a perturbation of the more general framework of undeformable plates. From a dynamic point of view, intraplate earthquakes in the oceanic plates are very valuable since they represent our principal clue to the state of stress of the lithosphere, and give us critical insight into the forces which may be driving the plate system. While seismic detection suffers in the remote areas of the ocean, the interpretation of the seismic data is made somewhat easier than on continents by the generally younger age of the ocean floor and the much simpler tectonic history of any oceanic province.

Our primary purpose in this lecture is to use intraplate earthquakes to derive information on the general state of stress of the oceanic plate and its mechanical properties. In this respect, we want to focus as much as possible on events associated with the average processes undergone by the plate as it ages, as opposed to earthquakes clearly associated with localized stress fields. Examples of the latter would be the 1975 Kalapana earthquake (the largest event ever recorded inside the Pacific plate; $M_s = 7.2$) which was due to a major volcanic intrusion, or seismicity associated with such large scale regions of diffuse deformation as the Ninetyeast Ridge area in the Indian Ocean, or the Caroline Plate in the Western Pacific. Thus, we shall adopt a rather restrictive view, and consider only as intraplate earthquakes events not controlled in their location or mechanism by present plate boundaries or by phenomena of a clearly extraordinary nature in the morphology of the oceanic plate.

2. Characteristics of intraplate earthquakes.

a. Seismic budget.

The existence of earthquakes inside the so-called stable blocks, including the Pacific Basin, was recognized by *Gutenberg and Richter* [1941], and their systematic study started with the work of *Sykes and Sbar* [1974]. As shown on Figure 8, intraplate seismicity is a common feature of all the oceanic basins in the world. Quantitative compilations of the level of seismicity have been given in the past five years, notably by *Bergman and Solomon* [1980], *Wiens and Stein* [1983] and *Okal* [1983, 1984]. In general, it has been found that the average rate of seismic moment release for intraplate earthquakes is on the order of 10^{26} dyn-cm/yr, nearly three orders of magnitude lower than the comparable rate for *inter*plate seismicity. *Wiens and Stein* [1983] have estimated the rate of seismic moment release per unit area to be $1 - 2 \times 10^{17}$ dyn-cm/(yr-km^2), and *Okal* [1984] has found a similar value for an extensive dataset of Pacific plate earthquakes, including historical events. Given the scarcity of intraplate seismicity it is difficult to estimate the rate of deformation of the plate; in one instance, it has been suggested to be 2% of the rate of accretion of the plate. These figures clearly justify treating intraplate seismicity as a perturbation of the general concept of rigid plates.

b. Magnitudes and the level of seismicity.

Intraplate regions are not immune to very large earthquakes. In oceanic areas, the largest events recorded have reached magnitude 7 along the continental margins of North America (Newfoundland, 1929; Baffin Bay, 1935), in the Northern (1983) and Southern (1947) Indian Ocean, and in the South Atlantic (1977). Most oceanic plates exhibit magnitude 6 seismicity; the Pacific plate is rather remarkable in this respect since the only magnitude 6 seismicity in this plate is concentrated along its fringes and in its extreme southern portion; an interpretation of this feature is given in Section 3.d. The seismicity of most intraplate oceanic areas below magnitude 4.5 is poorly known, because of sparse instrumentation.

Frequency-magnitude investigations of the intraplate seismicity of the oceans has usually failed to reveal significant departures from worldwide or interplate behavior. In particular, the clustering of seismic activity at low magnitudes typical of magmatic phenomena is observed only for distinct episodes of volcanism at a few sites with adequate seismic coverage, but is not a general feature of any of the major oceanic plates.

c. Detection and Location.

At least three factors contribute to substantially reducing our present detection capabilities in the oceanic environment, as compared to continental seismology. First, most oceanic stations are land-based i.e. located on islands, leaving most of the ocean basins uninstrumented. Second, because of swell-generated noise in the 1 – 5 s band, standard island stations operate at short-period gains as much as one order of magnitude below those of continental ones. Finally, higher seismic attenuation in oceanic structures further reduces detection capabilities. As a consequence, in most oceanic areas, the detection threshold is as high as $m_b = 4.7$. A notable exception to this situation is the French Polynesian network, where the use of narrow-band rejection filters to eliminate the swell-generated noise results in gains as high as 150,000 at 1 Hz, comparable to those achieved at standard continental stations [*Okal et al.*, 1980]. At the level

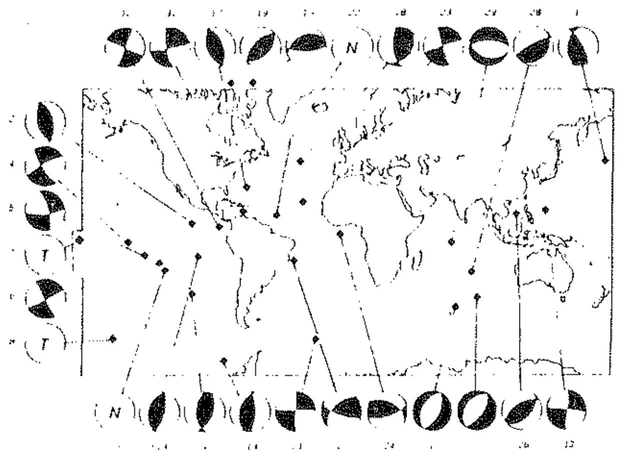
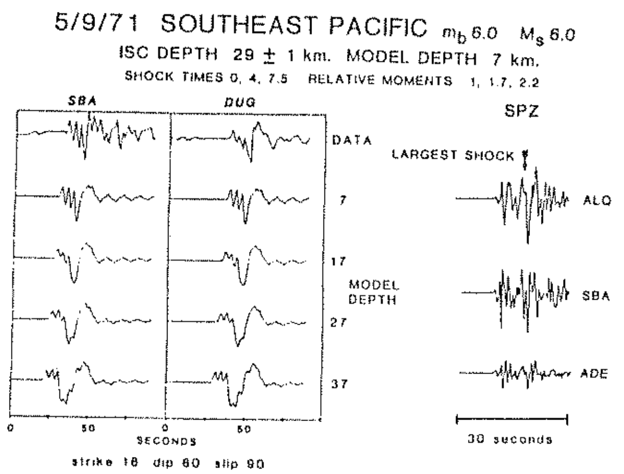
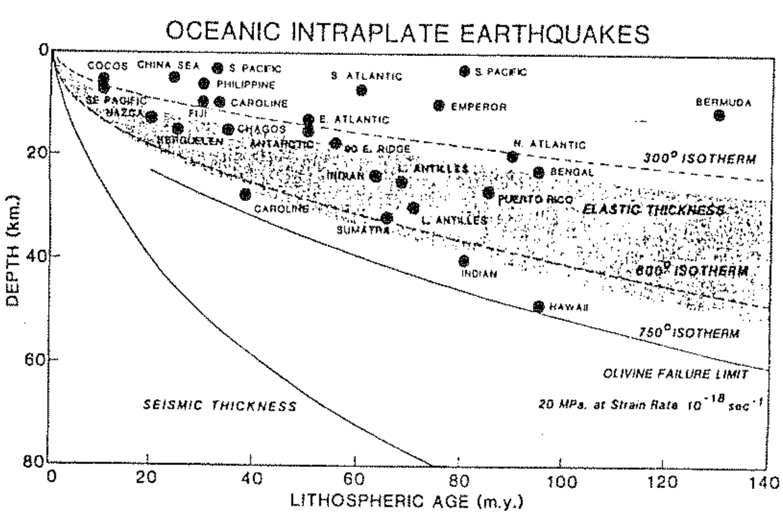


Fig. 8: Examples of oceanic intraplate earthquakes. Note their presence in all oceans, and the variety of focal mechanisms. N: unconstrained normal faulting; T: unconstrained thrust faulting [*Okal*, 1983].

Fig. 9: Depth of oceanic earthquakes [*Wiens and Stein*, 1983].
Right: Body-wave modelling as a function of depth. Note that an increase in the model depth of the event progressively adds to the complexity of the waveshape. When matched with the observed seismogram (top trace), this constrains the event's depth.



Bottom: Repartition of intraplate earthquake depths as a function of lithospheric age. Note the general thickening of the seismically active zone, and its good match of the elastic thickness of the plate. Also given are isotherms, and the 200 bar (20MPa)-strength line for a dry olivine rheology, at a strain rate of 10^{-18} s.



of $m_b = 5.0$, epicentral locations can be achieved with a reasonably good precision of ± 15 km.

d. Depth

The determination of the depth of intraplate oceanic earthquakes has long been a challenge. In particular, because of the general paucity of seismic stations in the ocean, the trade-off between depth and origin time leads to singularity in hypocentral inversions, and depths cannot be retrieved from arrival times alone. As late as 1965, the USGS proposed a depth of 143 km for a major Nazca plate event; it was relocated in 1971 to no deeper than 13 km. A large number of rather sophisticated techniques have been employed to constrain the depth of intraplate earthquakes. All suggest that oceanic foci are confined to the shallowest portions of the plate. One of the most elegant techniques uses the shape of teleseismic body waves to model the interaction between the direct seismic ray from the source to the station *(P)* and the rays reflected at the surface before traveling to the station *(pP* and *sP)*. The separation of these various contributions to the observed seismogram depends critically on depth, as illustrated on Figure 9.

Recently, *Wiens and Stein* [1983] used this technique to study 16 well-recorded intraplate earthquakes, and their results are presented on Figure 9*b*. They show that the maximum depth of seismicity increases with plate age to a maximum value of about 40 km. In particular, it is much shallower than the 80-km thickness of the "seismic" lithosphere, as measured from the dispersion of seismic surface waves (see Lecture 3). Exceptions would involve only cases involving continental margins, and earthquakes associated with intraplate volcanism (a priori excluded from our definition of intraplate events).

e. Focal Mechanisms

Once again, because of the lack of seismic stations at close distances from the epicenters, it is usually difficult to obtain first-motion constraints on focal mechanisms of intraplate oceanic earthquakes. Early studies, based entirely on teleseismic *P* waves, revealed only the general character of the solution (normal, strike-slip or thrust faulting). More complex methods, involving the modeling of body and surface waves, have allowed a more precise determination of the focal mechanism parameters of events with a magnitude of 5 or above.

Figure 8 shows that all kinds of focal mechanisms are represented in the oceanic lithosphere. However, there is clearly a majority of thrust and strike-slip events. All portions of the oceanic plates, regardless of age, can undergo thrust and strike-slip faulting. Normal faulting, on the other hand, is found only in very young areas of the plates: generally speaking, it is constrained to portions of the sea floor less than 15 Ma of age. Normal faulting is more frequent in the Indian Ocean, where it extends to 35 Ma. Normal faulting is unknown in older parts of the oceans.

In older lithosphere, a remarkable feature of the thrust and strike-slip seismicity is that it very frequently features a *common horizontal axis of compressional stress, oriented in the direction of motion of the plate away from the ridge*. The orientation of the horizontal tensional stresses for the normal faulting events in young lithosphere is more random, but in general, it *does not* point to the direction of motion away from the ridge.

3. Interpretation of intraplate earthquakes

a. Focal Mechanism: The Origin of the Released Stress

The discussion in this section proceeds under the assumption that intraplate earthquakes express the release of tectonic stress accumulated in the plate as the result of the large scale system of forces acting on the plates during their motion. In some cases (e.g., the readjustments following a major intraplate volcanic eruption), this assumption is clearly wrong, and this is why we reject these events *a priori*. In the ocean, *in situ* stress measurements are only starting to be taken, but in the continental environment, results from this technique have usually agreed with earthquake focal solutions, and thus upheld this assumption.

In the older ocean, the compressional stress directed away from the ridges has long been documented [e.g., *Forsyth*, 1973]. This is a general feature independent of the size of the plate, of whether or not it bears a continent, and of its absolute velocity with respect to the mantle. The origin of this stress is interpreted as follows: because young oceanic lithosphere is warm (if not hot), it is light, and buoyant, and thus elevated with respect to older, colder, denser lithosphere. Because it is higher, and despite isostatic compensation, it has a tendency to flow down and away from the ridge, creating a compressional stress in the adjoining older lithosphere. This system of forces acting from a young element of oceanic plate onto an older one, is known as *"Ridge-push"*, perhaps improperly because it does not originate exclusively at the ridge, but rather has to be integrated all over the plate. Its mathematical computation [e.g., *Turcotte and Schubert*, 1982; p. 287] predicts that it grows (grossly linearly) with the age of the plate, and creates stresses on the order of several hundred bars, sufficient to induce earthquake rupture.

Exceptions to this pattern involve only areas where strong deformation violating the basic rigidity of the plates is taking place (e.g. the Ninetyeast Ridge area), and isolated episodes, such as the Gilbert Islands earthquake swarm for which the common compressional axis was oriented perpendicular to ridge-push [*Lay and Okal*, 1983].

In the younger parts of the plate, ridge-push is rather weak, but may still explain the few occasional occurrences of thrust faulting. Several theories have been proposed to account for the more prominent normal faulting and its associated tensional horizontal stress, for example extrapolation of the tensional tectonics observed at the MOR's, or thermal stresses induced by the shrinking of the plate as it cools away from the ridge. The observation that the tensional stress is preferably oriented parallel rather than perpendicular to the ridge, clearly favors the second model, and is in general agreement with other data in support of a very limited (50 km or so) domain of extensional tectonics at the MOR. However, the fact that all three kinds of earthquake mechanisms can co-exist within very small distances (e.g., 65 km in the Pacific Ocean) clearly indicates that the stress release in young oceanic plates is controlled by small-scale effects that are not presently fully understood.

b. Depth of Seismicity: Inferences about the Rheology of the Plate

The observation that the maximum depth of seismicity increases with the age of the plate (see Figure 9*b*) leads naturally to the idea that it is controlled by temperature, which is known to similarly vary with age. Numerical modelling of the temperature field of the oceanic plate as a function of age indicates that the region of seismicity is bounded by an isotherm ranging from 600°C to 750°C. As such, the thickness of the seismically active region can be compared to the flexural elastic thickness of the plate. Specifically, earthquakes, which represent rock fracture in the brittle mode, can occur only in the domain of brittle failure of the rock. The extent of the brittle and ductile domains, as a function of depth and temperature, depends on a number of parameters, including pressure, pore pressure (of water, if the rock is wet), and strain rate. For strain rates of 10^{-18} to 10^{-15} s^{-1}, and a dry olivine rheology, a strength of 200 to 1000 bars is predicted at the 750°C isotherm. Weaker rheologies, involving for example wet olivine, with pore pressures approximately 70% lithostatic, would require extremely fast strain rates (10^{-10} s^{-1}) to account for the observation of deep seismicity [*Wiens and Stein*, 1983].

c. Location: The Preferential (?) Siting of Seismicity on the Plate.

Finally, we want to address the problem of where and how release of the tectonic stress (e.g., ridge-push) takes place. This is prompted by the frequent observation of self-repeating earthquakes (either in the form of swarms or of isolated events) at the same epicenters, suggesting locally a particular "weakness" of the plate.

On continents where geomorphological features are accessible for mapping, a number of observations have led to the concept of zones of weakness, along which preferential tectonic activity, such as intraplate seismicity, and the intrusion of kimberlites, are emplaced [*Sykes*, 1978]. A number of investigators [e.g., *Bergman and Solomon*, 1980; *Okal*, 1984] have thus attempted to correlate oceanic intraplate epicenters with prominent bathymetric features, in the hope of recognizing preferential lines of weakness in the plate. This endeavor is made very difficult by the extremely sparse, and occasionally biased, bathymetric coverage of a large fraction of the world's oceans. Recently, the use of satellite-derived bathymetry, such as the *SEASAT* dataset, has helped alleviate this situation, but it provides mostly large-scale long-wavelength information and exclusively for uncompensated features. Results from these studies are rather mixed: it has been suggested that the healed portions of fracture zones, outside their active transform segments, may be areas of preferential seismicity (e.g., for the 1978 Bermuda earthquake). Other lines of weakness could include former plate boundaries (e.g., the Ninetyeast Ridge or a line in the panhandle of the Antarctic plate), the traces of hotspots (e.g. "Region C" in the Pacific, where rift propagation took place in the Miocene), and lines of age discontinuity in the lithosphere, brought about, for example, by changes in the spreading pattern such as ridge jumps (e.g., in the southeastern Pacific). However, in some cases, detailed bathymetric investigations of areas of concentrated seismicity have failed to reveal any particularly anomalous feature in the sea floor. It is clear that we do not always understand the siting of seismicity on the plate.

d. The Particular Vulnerability of most of the Pacific Plate.

In the Southeastern Pacific Ocean, a major readjustment of the spreading pattern took place between 25 and 18 Ma ago. The Farallon Ridge which formed the Eastern boundary of the Paciﬁic plate became inactive, and spreading was transferred to a new ridge, the present-day East Pacific Rise, approximately 500 km to the West; this ridge *jump* was accompanied by a clockwise *reorientation* of about 40°: as a result, the fossil fracture zones of the old system strike about 250°, while the present direction of spreading away from the ridge is 290°. This discrepancy between the morphological directions on the old plate and the direction of the tectonic stress applied to it creates a situation of favorable rupture along pre-existing blocks, in the strike-slip geometry. In the absence of reorientation, the same stresses can be taken-up along existing faults only in the less favorable thrust mode (see Figure 10). *Okal* [1984] has proposed this model to explain the prominence of strike-slip faulting and the absence of magnitude 6 seismicity in the part of the Pacific plate which underwent the reorientation. Significantly, these characteristics are absent from the other parts of the plate.

e. Finally, we want to stress that in many areas of the ocean where seismic coverage is poor, the interpretation of isolated seismic events as expressing a large scale tectonic

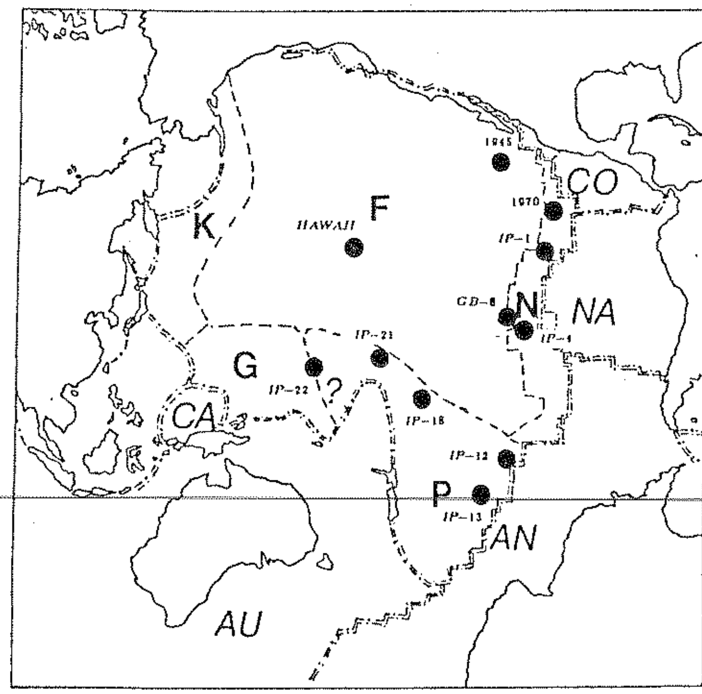
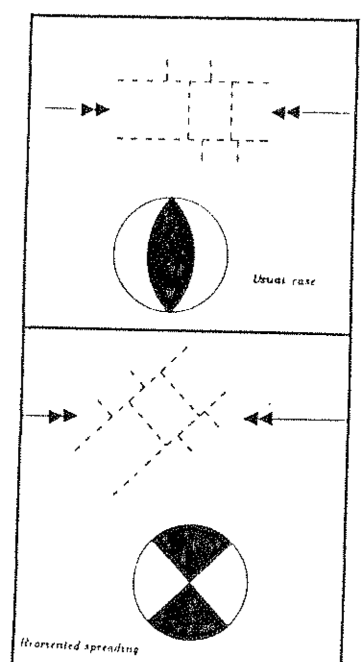


Fig. 10: Vulnerability of the bulk of the Pacific plate [*Okal*, 1984]. The top diagram shows that, except for the area of Hawaii, the part of the Pacific plate generated at the old Farallon ridge (and labeled "F") is deprived of seismicity above magnitude 6. This particular ridge underwent a reorientation of spreading about 20 Ma ago, resulting in the vulnerable geometry of stresses and lines of faulting shown on the bottom diagram. The fringes of the Pacific plate, which were not involved in the change of spreading direction, are less vulnerable, and rupture at higher magnitudes.



process must often remain tentative: specifically, such an isolated event could represent the "tip of the iceberg", namely the only recorded earthquake in a more important, but unsuspected swarm, which may, for example, have had a volcanic origin. Examples of volcanic swarms which were detected only by the local network abond in French Polynesia, and an example exists at Deception Island (Antarctica) of a volcanic eruption, occurring on land and therefore well documented, but from which only one major earthquake was recorded, six months later, corresponding to tectonic readjustment of the volcanic edifice. Had the volcanic eruption been concealed underwater, this earthquake might have been interpreted in a totally wrongful context.

LECTURE 5 - ACTIVE TECTONISM ON THE OCEANIC PLATES

1. Large scale volcanism: hotspots

a. Introduction

The linearity of islands chains, notably in the Pacific Ocean, has long been recognized. Indeed, Wegener had already commented on this feature in his book on Continental Drift, where he mentions that islands chains are of two types: a first type parallel to continental masses and perpendicular to their drift direction (we would now call them *island arcs at subduction zones*), and a second type aligned perpendicular to the first one, and located in the middle of the oceans. The Hawaiian Islands are a primary example of the latter.

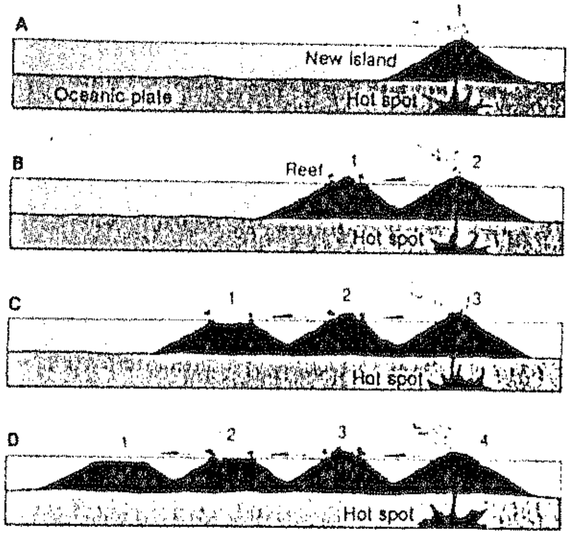
Following systematic dating of igneous rocks from oceanic islands, *J. Tuzo Wilson* [1963] made the fundamental observation of a linear correlation between the age of volcanism and distance along the oceanic chain. Furthermore, the age of an oceanic island is usually different from that of the adjoining ocean floor: for example, Kilauea on the big island of Hawaii is still an active volcano, although it sits on 75 Ma old lithosphere. Wilson then proposed to interpret his results using the so-called **"hotspot"** model (see Figure 11), in which the oceanic plate glides over a magma source *fixed with respect to the mantle*. At any time, only the island located right on top of the "hotspot" is active; as time goes by, it is pushed away by the motion of the plate, becomes inactive, is eroded and eventually becomes an atoll, and later a submerged "guyot". This model explains the linear relation between age and distance and the asymmetry of the chain.

"The islands are in fact arranged like plumes of smoke [....] carried downwind from their sources" (J. Tuzo Wilson).

b. Properties of hotspot chains

The typical example of a hotspot chain is the Hawaiian Islands. Thus we will detail its properties:

- The big island of Hawaii holds two active volcanoes: Mauna Loa and Kilauea, as well as three extremely recent ones: Mauna Kea, Haleakala and Kohala.
- The next island, Maui, is dormant and was last active in 1790.
- Volcanism as young as 40,000 years is known on Oahu (the "main" island where Honolulu is located). These rocks are known as the "Honolulu series". Most rocks on Oahu are about 4 Ma in age.
- Northwest of Kauai, the islands have been eroded to sea level, and only *atolls* remain. Rocks drilled from the last island in the chain, Midway Atoll, have been dated to 19 Ma.



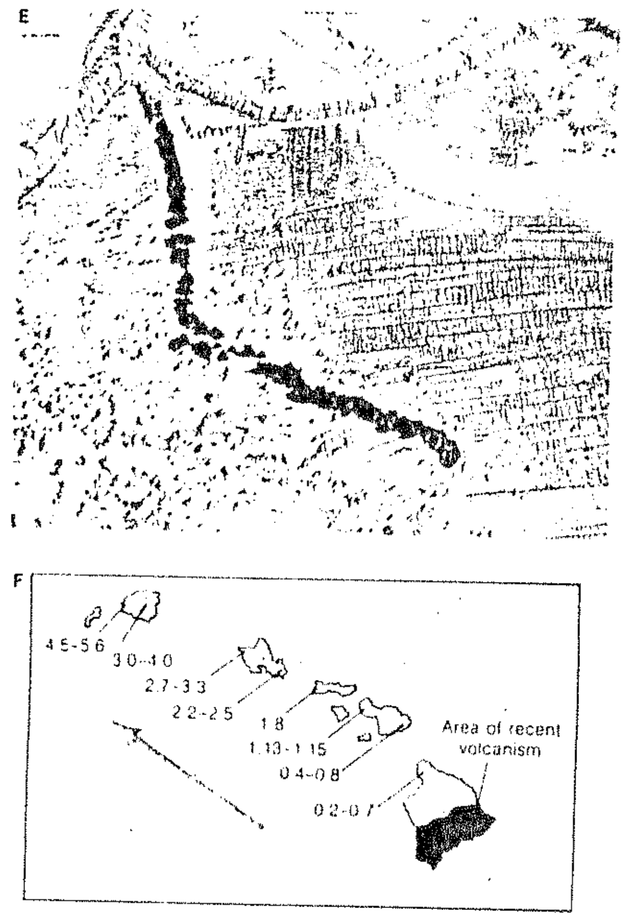


Figure 19.13 The origin of linear chains of volcanic islands and seamounts is explained by the plate tectonic theory. **(A)** A volcanic island or seamount is built up by extrusions from a fixed hot spot, or source of magma, in the mantle. **(B)** As the plate moves, the volcano moves away from the source of magma and becomes dormant. The surface of the island can then be eroded to sea level, and reefs can grow to form an atoll. A new island is formed over the hot spot. **(C)** Continued plate movement produces a chain of islands. **(D)** The islands of the chain are progressively older away from the hot spot. **(E)** An abrupt change in the direction of plate movement is indicated by a change in the direction of a chain of islands. The Emperor seamount chain began to form more than 40 million years ago, when the Pacific plate was moving northward. About 25 million years ago, the plate moved northwestward and started to form the Midway-Hawaiian chain. **(F)** The Hawaiian Islands are progressively older away from the area of recent volcanic activity. The numbers refer to the ages, in millions of years, of the volcanic basalts.

Fig. 11: Sketches of the "hotspot" model [*Hamblin*, 1982].

Beyond Midway, only *seamounts* remain, partly because the increasing latitude prevents the growth of coral to offset the erosion, but the chain can be recognized for another 1200 km.

Southeast of Hawaii, the youngest volcano, a seamount known as Loihi, topping at 980 m b.s.l. has been recognized as an active seismic center and is interpreted as the "next island", presently in a growing stage.

An obvious character of hotspot volcanism exemplified by the hawaiian volcanoes is the low viscosity of its lavas. The volcanoes make up huge *shields*, with a typical base angle of about 15°. Also, explosive eruptions are very rare, and replaced by nearly continuous activity. Petrological investigations confirm these results: most hawaiian lavas are *tholeiites* fairly rich in silica with respect to alkalis.

In the extreme Northwest Pacific, the Hawaiian chain (which by then is composed of seamounts) takes up a sharp bend, and runs North-Northwest rather than West-Northwest. The northern branch is known as the Emperor seamount chain, and can be followed all the way until it subducts into the Kamchatka-Aleutian corner. This feature is interpreted as resulting from a change of the direction of motion of the Pacific plate, with respect to the mantle and thus to the hotspot, around 45 Ma. It is believed that this may be a secondary effect of the separation of Australia from Antarctica, which started 53 Ma ago.

d. Other hotspot chains

Other island chains in the Pacific, strikingly parallel to the Hawaiian Islands, include:

The Society Islands: (from young to old) Mehetia, Tahiti, Moorea, Huahine, Raiatea, Bora-Bora.

The Cook-Austral Chain: Macdonald, Rapa, Tubuai, Rurutu and the Cook Islands.

The Pitcairn Chain: Pitcairn, Mangareva (Gambier), Mururoa, Duke of Gloucester.

Systematic studies of the age-distance relationships have shown that the relative motions of a rigid Pacific plate with respect to the various hotspots are compatible with the idea of the hotspots being fixed with respect to each other. This is usually upheld to a precision on the order of a few mm/yr, as to compared to the 10 cm/yr typical of the motion of the plate with respect to one hotspot. It can also be interpreted as further proof of the *rigid* character of the plates.

Many authors have also noticed that some smaller island groups in the Pacific are oriented parallel to the Emperor seamount chain (e.g. the Line Islands), and have sought to prolong the above-mentioned chains. This prolongation is made necessary by the following constraints: If the rigid Pacific plate has changed direction with respect to the Hawaiian hotspot 45 Ma ago, it must have changed direction with respect to the other ones as well; furthermore, evidence along the Hawaii-Emperor chain suggests that hotspots are very long-lived features. *We should be extremely reluctant to having them turned on and off at will.* As a consequence, the other Pacific chains should also extend past a bend, contemporary with the Hawaii-Emperor bend. Efforts at understanding the formation of islands in the Pacific Ocean with a limited number of hotspots are still being carried out.

Hotspot chains in other oceans include (from young to old): Réunion-Mauritius; Heard-Kerguelen; and Cape Verde. If the hotspot is presently located under a spreading ridge, it leaves *two* wakes (one in each plate). This is the case for Iceland and the Iceland-Faeroe plateau in the European plate, the Iceland-Greenland plateau in the North American plate, and also for Tristan da Cunha, and the Walvis and Rio Grande Ridges in the South Atlantic.

In some cases, hotspots are presently located under continental masses. Examples would include Mount Cameroon in Africa and Mount Erebus in Antarctica. Another example is Yellowstone, a dormant volcano whose wake is the Snake River plain.

2. Origin of hotspot volcanism

a. Isotopic composition

Systematic studies of the isotopic composition of igneous rocks from oceanic islands have shown that their geochemistry is different from that of Mid-Oceanic Ridge Basalts [MORB's]. Specifically, oceanic island basalts are, relative to MORB's, richer in alkalis such as Rubidium [Rb], and thus described as coming from an *undepleted* [in alkalis] source. This source of mid-plate volcanism is also found to be depleted, relative to the MORB reservoir, in the Rare Earth Element Samarium [Sm]. The nuclear decay of ^{87}Rb contributes to an enrichment of the isotopic ratio ^{87}Sr/^{86}Sr, and similarly the decay of ^{147}Sm contributes to the isotopic ratio ^{143}Nd/^{144}Nd. From the analysis of these ratios in different minerals of the same igneous rock sample, it is possible to compute the value the isotopic ratios had at the time of formation of the rock. This gives information on the geochemistry of the magma from which the rock was derived. The fact that hotspot reservoirs are *undepleted*, i.e. have higher ^{87}Sr/^{86}Sr, and lower ^{143}Nd/^{144}Nd is generally interpreted as meaning that they do not participate in the MORB cycle, and thus probably originate at depths *greater* than the depth of continental differentiation.

It is important to realize that these isotopic properties are far more characteristic of hotspot material than the mere difference in age between island and ocean floor: for example, if the hotspot is on the ridge, there will be no age difference, but still the different chemistry will serve proof that the origin of the island is different from that of the rest of the ridge.

b. Deep Roots ?

That leaves open the question of the depth of the hotspot magma chamber, and more generally speaking of the origin of the hotspots. Not surprisingly, Seismology has been used (and misused) in efforts to map anomalies below hotspots, so as to gain insight into their depth extent.

Around the beginning of the 1970's, hotspots were usually given a very deep structure, following claims that seismic rays under Hawaii were deviated by the presence of lateral heterogeneities, as deep as the core-mantle boundary. Later, a number of investigators, including *Okal and Kuster* [1975] showed that the heterogeneities responsible for anomalies in the propagation of the rays were actually much shallower, and located under the receiving network.

The idea that hotspots could be anchored all the way to the core-mantle boundary was used to propose that the upwelling mechanism involved in hotspots, especially those located on ridges, was an active agent in the plate-driving mechanism. Both of these ideas are to a great extent abandoned nowadays; their main problem was that the seismological evidence on which they were based was at best shaky. The maximum depth at which heterogeneity has been reported under a hotspot is about 250 km below Yellowstone, 300 km under Iceland.

c. Island Formation

One of the most striking features of island chains in the middle of oceanic plates is their discrete character: volcanism takes the form of distinct, individual islands, separated by a regular oceanic column (down to 5000 m in between islands of the Hawaiian chain), rather than a continuous volcanic ridge. This situation has been modeled by the diapirism of a light, low-viscosity fluid inside a matrix made of a heavier, higher-viscosity material. Theoretical investigations indicate that a plume will form; in other words, the rise of the lighter material will occur along just a few pipes rather than homogeneously; additionally, in the presence of a very high-viscosity lid moving laterally above the two-fluid system, it is found that the pipes can *bend* and remain stable up to a certain angle. After that, a new pipe branches out and starts bending until it too breaks off [*Skilbeck and Whitehead*, 1978]. This clearly results in several, discrete islands rather than a continuous ridge. However, in the case of a plate with zero thickness, such as when the hotspot is on-ridge, the argument for bending the pipe disappears, and a continuous ridge is formed. This is the case for the Iceland-Faeroe plateau or the Walvis Ridge.

3. "Non-hotspot" volcanism

The simple ideas behind the hotspot theory cannot explain all intraplate volcanism observed in the Pacific Ocean. In particular, in the Austral Islands several episodes of volcanism separated by as much as 17 Ma are documented on individual islands. This goes against the idea that only one island is active at any time in a hotspot system. Similarly, in the Line Islands, recent radiometric dating using $^{40}Ar/^{39}Ar$ shows that sites several hundred km apart must have been active concurrently during the Late Cretaceous. In order to explain such volcanic episodes in the simple hotspot theory, one would have to multiply the number of hotspots; each of the new hotspots so involved at any one time, must be extrapolated both in the past in the present, or turned on and off to satisfy local conditions. This approach goes against the simple assumptions of the hotspot model.

Finally, the ocean floor is covered with seamounts, with [presumably] many more still to be discovered, and it would be impossible to model all of them as traces of long-lasting hotspots, since that would require a systematic pattern of linear chains. There is no fundamental difference between the volcanic properties of the large island chains, such as Hawaii, and those of the small volcanoes found on the ocean floor. This observation has led *Batiza* [1982] to the concept of the "non-hotspot" volcanism being a failed, or stillborn, expression of an episode comparable to hotspot diapirism. The major difference between the two is then the persistence of the active diapir over periods of time large with respect to the lifetime of one individual volcanic edifice (~1 Ma), most probably due to a question of size of the diapir. This model partially alleviate the problem of switching hotspots on and off, while preserving the hotspot concept, which clearly works well for the major edifices, such as Hawaii and the Society Islands.

4. The Evolving Plates: stability and relationship of ridge and hotspot systems

The system of plate boundaries under which the Plate Tectonics cycle operates is by no means frozen, and slowly evolves with time. It is usually believed that the major components of the plate-driving forces include the gravitational pull of subducting slabs. As the age of subducting material varies, these forces themselves change, and a different kinematic pattern may result. One particular case is that of a fast-consuming plate, whose ridge eventually reaches the subduction zone: because of the buoyancy of the young material at the ridge, subduction is expected to cease, and also the ridge becomes inactive, resulting in a change in the driving system.

Other patterns in the evolution of the ridge system include *rift propagation:* As suggested by *Hey* [1977], the position of a Transform Fault offset along a ridge may become unstable, and gradually propagate down the ridge, resulting in an increase of one of the active ridge segments at the expense of the other. A number of current research projects aim at identifying the agents responsible for this mechanism, which gives rise to active tectonism on the ocean floor, recognizable as "V"-shaped structures with complex bathymetry and magnetics.

An imperfect and incomplete form of rift propagation can lead to simultaneous activity along two parallel ridge segments, and thus to the development of a small "platelet", which may be only a transient feature. Such is believed to be the present-day situation at the Easter plate in the Southeastern Pacific Ocean. Although such platelets can be ignored when describing large-scale plate kinematics, their correct interpretation gives insight into some transient dynamic phenomena.

Eventually, when a ridge fails, it remains as a fossil feature in the bathymetric and magnetic structure of the ocean floor. The static interpretation of these features can be given in terms of a ridge jump. Important episodes of ridge jumping occurred in the Southeastern Pacific, in the Indian Ocean along the Ninetyeast Ridge, and in the Atlantic Ocean across Greenland. The time it takes for a ridge to jump can vary, and has been estimated anywhere from a few million years in the Pacific to 20 Ma in the North Atlantic.

One of the major successes of the hotspot models resides in their stability in time: hotspots have been routinely traced for more than 100 Ma in the Pacific and *Chase and Sprowl* [1983] have proposed to relate their origin to ancient geometries of the Plate Tectonics cycle going back some 200 Ma. On a shorter time scale, hotspots tracks are found to be particularly robust to the presence of tectonic features such as ocean-continent transitions, and the crossing of an inactive fracture zone system.

On a time scale not exceeding a few 10^7 years, it has been proposed that hotspots may affect the kinematics of plate boundaries, either by contributing to the initial break-up of a plate, or by "trapping" a neighboring ridge. This appears to have been the case in Iceland, which has trapped the Reykjanes Ridge over the past 20 Ma through a series of ridge jumps. However, in other situations, such as Kerguelen, and the Northern Tuamotu Islands in the Pacific, the hotspot clearly has left the ridge, after possibly providing the mechanism for a rift propagation lasting about 11 Ma [*Okal and Cazenave*, 1984]. Conversely, in a related area of the Pacific, these authors have proposed that a young fossil fracture zone may have deviated (by no more than 15°) the surficial expression of a hotspot. Also, *Morgan* [1978] has proposed the existence of pipes leaking hotspot magma into a nearby MOR system. There emerges a picture of the possibility of limited interaction between the ridge system and the hotspots, and the distances (~500 km) over which this interaction is observed may be representative of the size of the thermal and mechanical anomaly termed a "hotspot".

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