Plate Tectonics and Hotspots: The Third Dimension

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High-resolution seismic tomographic models of the upper mantle provide powerful new constraints on theories of plate tectonics and hotspots. Midocean ridges have extremely low seismic velocities to a depth of 100 kilometers. These low velocities imply partial melting. At greater depths, low-velocity and high-velocity anomalies record, respectively, previous positions of migrating ridges and trenches. Extensional, rifting, and hotspot regions have deep (>200 kilometers) low-velocity anomalies. The upper mantle is characterized by vast domains of high temperature rather than small regions surrounding hotspots; the asthenosphere is not homogeneous or isothermal. Extensive magmatism requires a combination of hot upper mantle and suitable lithospheric conditions. High-velocity regions of the upper 200 kilometers of the mantle correlate with Archean cratons.

Seismic tomography is a new but rapidly developing method for imaging Earth's interior (*1–4*). Early results had limited resolution, but important problems regarding the geoid and the large-scale structure of mantle convection have been solved (*5*). Seismic methods image the interior with a resolution and coverage that are limited only by wavelength and the distribution of earthquakes and seismic stations. Major improvements have recently been made as a result of great increases in the surface wave data. It is now possible to use a block parameterization rather than the less flexible spherical harmonic representation (*4*) and to resolve such features as midocean ridges, cratons, and dead slabs. With these high-resolution three-dimensional images of the mantle, we address some of the current issues of plate tectonics and hotspots.

The directions and speeds of plate motions are generally attributed to boundary and plate forces such as slab pull and ridge push (*6*). Midocean ridges are passive reactions to plate forces, not the tops of convection cells. On the other hand, hotspots are viewed as the tops of active upwellings. Hotspots are long-lived centers of extensive volcanism, such as Hawaii, Iceland, and Yellowstone, and they account for between 5 and 10% of the heat and magma expelled from the mantle (*7, 8*). They are generally attributed to deep-mantle plumes and a scale of convection different from that of plates. In most models of plate tectonics and mantle petrology the upper mantle is viewed as well-mixed, homogeneous, and isothermal except near hotspots. The style of sublithospheric mantle convection has been uncertain because of the filtering effect of the plates and the strong contribution to topography and gravity by crustal variations. With tomography we can now see beneath the plates.

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Images

The tomographic results are presented as maps of shear wave velocity at depths from 38.3 to 410 km (Figs. 1 through 5). The faster regions, generally corresponding to cold or refractory material, are given in blue while the slow, generally hot, regions are dark orange (*9*). The "plates," of "plate tectonics," average about 100 km in thickness. The oldest continental plates involve Precambrian cratons and are much thicker. They show up as dark blue areas (+4%) in the maps for the shallow layers. Midocean ridges, tectonically active regions, and volcanic regions have low velocity (Figs. 1 to 3). The correlation with present-day tectonics degrades with depth (Figs. 4 and 5). In principle, the planform of mantle convection should become evident once we get well below the lithosphere and the plate-entrained flow at the top of the mantle.

Instead of seeing well-defined convection cells, including hot upwellings, we see large domains of low- and high-velocity mantle (Figs. 1 through 5). Inside the very large low-velocity anomalies (VLVAs) are even slower regions that may represent hot, volatile-rich upwellings or hot cells. There are high-velocity anomalies (HVAs) at all depths. The deeper ones (Fig. 5) possibly represent cold oceanic lithosphere that has sunk into the mantle. The seismic velocity variations at depths below about 200 km seem to have more to do with past, than with present, plate tectonics. This is expected because continents and plate boundaries are quite mobile compared to convection cells in the mantle; also, cold-subducted lithosphere has a long thermal time constant.

Figures 1 and 2 show the shear velocities near the top of the mantle. Note the spectacular "ring of fire" around the Pacific and the low velocities associated with ridges and deep-sea trenches. Old crust and tectonically stable regions are underlain by fast mantle due to a combination of low temperatures and a refractory mineralogy (*10–13*). The extremely low velocities in convergent regions (for example, western Pacific trenches), particularly those with backarc spreading, may be due to dehydration of subducted crust and volatile-fluxed melting of the overlying mantle. The lowest velocity regions are probably partially molten (*11*).

Oceanic Mantle

The most impressive features of the tomographic maps are the broad low-velocity anomalies (LVAs) in the Pacific and Indian oceans. Clearly, these oceans are different from the newly opened oceans, the Atlantic and Arctic. The Pacific is particularly slow from the East Pacific Rise (EPR) toward the northwest, generally in the spreading direction. Hotspots (black squares) appear to be randomly distributed in broad low-velocity regions. Most of the world's oceanic crust is formed at the EPR, and most of the world's hotspot volcanism (Hawaii, Polynesia) occurs on the Pacific plate, downstream (that is, in the spreading direction) from the EPR. Many spreading-center LVAs extend only to 100 km, but some in the Pacific and Indian oceans extend deeper than 300 km. Most oceanic hotspots and backarc basins have LVAs extending to about 200 km. Note the contrast between the slow Pacific and the fast Eurasian and North American hemispheres. There is a broad LVA in the central Pacific, which may be related to extensive igneous activity in the Cretaceous (*12, 13*). The large plateaus in the western Pacific formed during a period of global plate reorganization, continental breakup, and rapid spreading. The extensive oceanic ridge and hotspot magmatism in the Pacific at present is probably a result of these large areas of hot mantle.

Midoceanic ridges have very low seismic velocities in the far North and far South Atlantic, eastern Pacific, southeastern Indian Ridge, around triple junctions, and near some hotspots. The Mid-Atlantic Ridge (MAR) has LVAs at 30° and 40°N. The former may be the source of the New England Seamount–White Mountain track (*14*). The latter is near the Azores triple junction. Regions of extension (Rhinegraben, Kenya Rift, southwestern United States, Lake Baikal) have shallow LVAs. Most of the Atlantic and Indian ocean ridges faithfully follow the lowest velocity regions in the shallow mantle (Figs. 1 and 2). The lack of correlation at greater depths might suggest that ridges are passive, with hot material upwelling into the space made available by plate spreading. Melting occurs during decompression, and this causes very low shear velocities. It must be recognized, however, that midocean ridges are mobile and can drift away from the spreading-induced upwellings. The Pacific ridge systems are embedded in very broad low-velocity areas (VLVAs). Note the discontinuity near the Australian-Antarctic ridge. This is an anomalous ridge in depth and geochemistry (*15*). It appears that the eastern part is being fed from the Pacific side, consistent with the geochemistry.

Backarc basins (western Pacific) and sites of recent subduction are among the slowest uppermost mantle regions. The slowest areas are western North America, New Zealand, southeastern Asia, the Philippine Sea, eastern Pacific, and northeastern Australia–Coral Sea. Theories that associate all hot regions of the shallow mantle with deep mantle plumes must contend with these LVAs. The older parts of the oceans have fast shallow mantle, indicating cold and thick lithosphere. In most models of plate tectonics, bathymetry is due to lithospheric thickening over a homogeneous, isothermal mantle with perhaps a few interruptions by hotspot swells, the tops of deep mantle plumes (*16, 17*). The tomography indicates that this is too simple a view.

Hotspots

Hotspots are those volcanic regions that are thought to be distinct from island arc or midocean ridge magmatism. Some hotspots are at the end of a trail of islands and have been attributed to deep mantle plumes

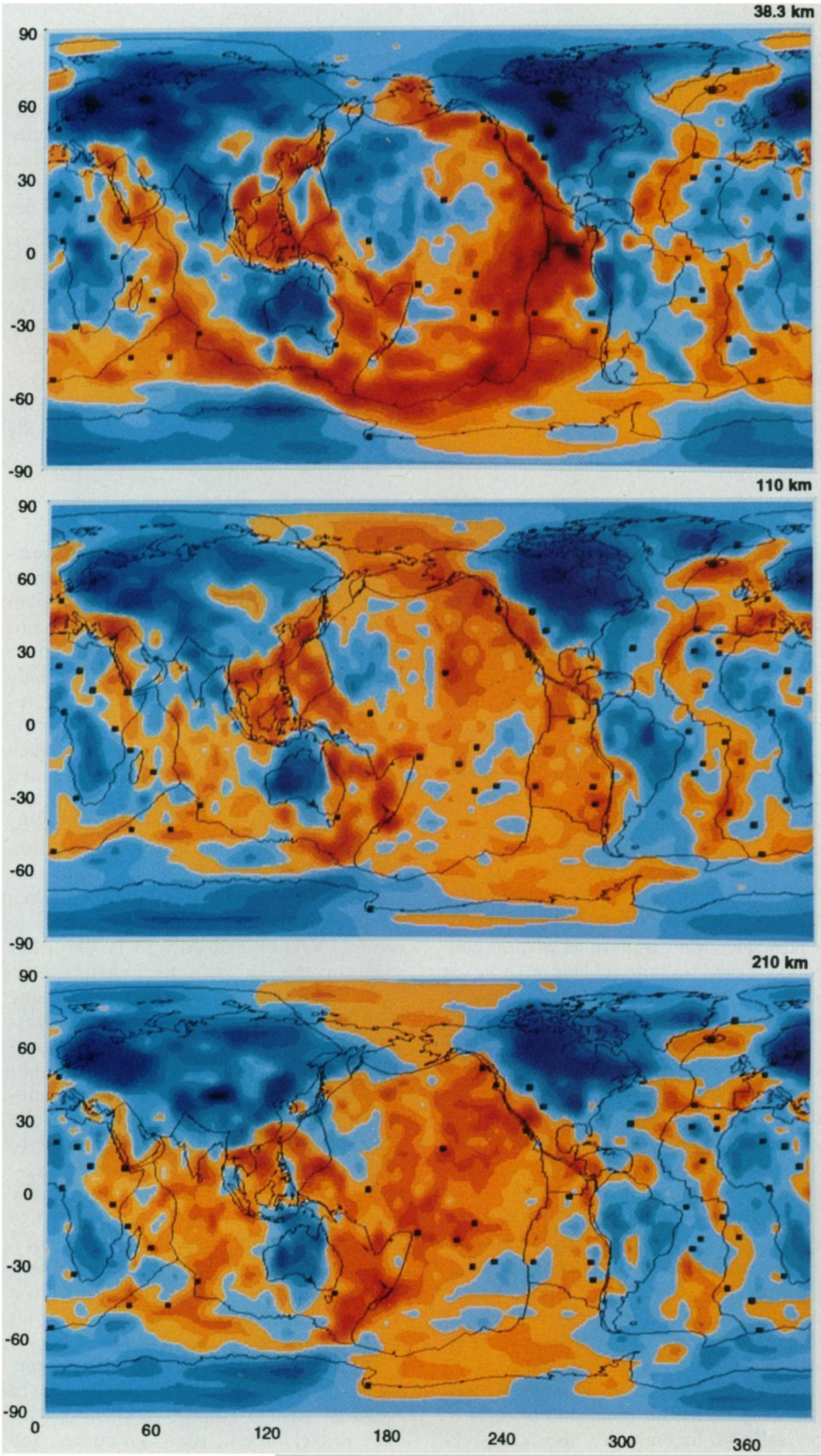

Fig. 1 (top). Shear velocity at a depth of 38.3 km from the model of Zhang and Tanimoto (*4*). Continental and plate boundaries are shown; hotspots are black dots. Blue regions have high seismic velocities. Orange regions have low velocities (hot). Range of velocity is ±5%. The dark blue regions are stable cratons or old oceanic lithosphere. The slowest regions are young oceanic lithosphere and tectonically active regions. The hottest regions follow the midoceanic ridges and subduction zones (backarc basins). **Fig. 2 (middle).** Shear velocity at a depth of 110 km. Total range in velocity is ±4.5%. Note the sinuous LVA following the Atlantic and Indian ocean ridges. Continental shields are very fast. **Fig. 3 (bottom).** Shear velocity at a depth of 210 km. Total range is ±4%. Note the sinuous LVAs in the Atlantic and Indian oceans, which are offset from the current ridges. There are VLVAs in the Pacific and Indian oceans, especially from New Zealand to California.

(*18*). Plumes may represent the main mantle upwellings (*17, 18*) or secondary, small-scale convection (*8, 19–21*). Although hotspots originate below the moving plates, their actual depth of origin is unconstrained. Bathymetry, geoid, and heat flow data across hotspots can be satisfied by thermal perturbations confined to the upper 200 km of the mantle (*22*).

Most of the Atlantic and Indian ocean hotspots are embedded in hot, slow shallow bands related to spreading ridges, past or present. Hotspots favor near-ridge locations. If the migration of ridges is taken into account, the close association of hotspots with ridges, present and recent, is even more impressive. Many hotspots are located in VLVAs (Iceland, Hawaii, Azores), but the sizes, shapes, and trends of these VLVAs are not related to plate motions. There are other similar VLVAs that are devoid of hotspots. VLVAs may provide massive outpourings of basalt, such as continental flood basalts (CFBs) or oceanic plateau basalts, when plate reorganization throws the overlying lithosphere into relative extension. Tomography shows that both ridges and hotspots generally overlie large regions of hot upper mantle. Few hotspots have the mushroom-shaped plume-head LVAs predicted by plume theories. Hotspots are not centrally located in VLVAs, nor is hottest mantle always associated with hotspots. The location of hotspots and large igneous provinces near present and past positions of spreading ridges, cratons, and other lithospheric discontinuities suggests a strong control by the lithosphere.

The most active hotspots and ridges are associated with broad LVAs, presumably large hot areas of the upper mantle. Most hotspots are in LVAs but a map of hotspots does not provide a particularly good description of hot regions of the mantle or, therefore, of mantle convection (Figs. 2 and 3). Most hotspots lie near the edges of LVAs or in high-velocity anomalies. Many VLVAs are not near hotspots. It is likely that a combination of hot mantle plus appropriate lithospheric stress conditions is required in order for the upwellings that we call hotspots to occur. For example, the absence of hotspots above VLVAs in the northern Indian Ocean is likely due to lithospheric compression. There are LVAs in the wakes of the Americas, India, Australia, and Greenland. These are thick-lithosphere (Fig. 1), low-geoid regions. Hotspots may not be narrow pipelike features in the mantle but rather the focused effect of upwelling triggered from above by extensional strains or discontinuities in the lithosphere.

Many hotspots are underlain by colder than average mantle below depths of 200 to 300 km (Figs. 3 and 4). These include St. Helena, Tristan da Cunha, Iceland, Easter, Bouvet, and Yellowstone. If we look at fast velocities in the transition region (Fig. 5), we can add Azores, Ascension, Galápagos, Marquesas, Crozet, and the Carolines. Almost all of these hotspots are close to lithospheric discontinuities such as triple junctions, fracture zones, or rift zones. Thus, there is still no geophysical evidence that demands deep thermal perturbations beneath hotspots (*22*). Many hotspots are hot lines, or erupt for long periods of time after leaving the "plume," or do not exhibit the simple age progressions, or are up to 1000 km away from a conjectured plume, or are definitely not fixed (*23*). If hot mantle has the dimensions and shapes of the seismic LVAs, then these observations, combined with lithospheric control of exit points, can be understood. Hotspots should appear to wander or jump or turn on or off when plate motions change. If plate reorganization causes ridges or triple junctions to open up in the Pacific VLVAs there will probably be massive outpourings of basalt, as there was in the Cretaceous, without the importation of a deep mantle plume. Hot cells may have been mistaken for narrow plumes.

Some hotspots that are embedded in

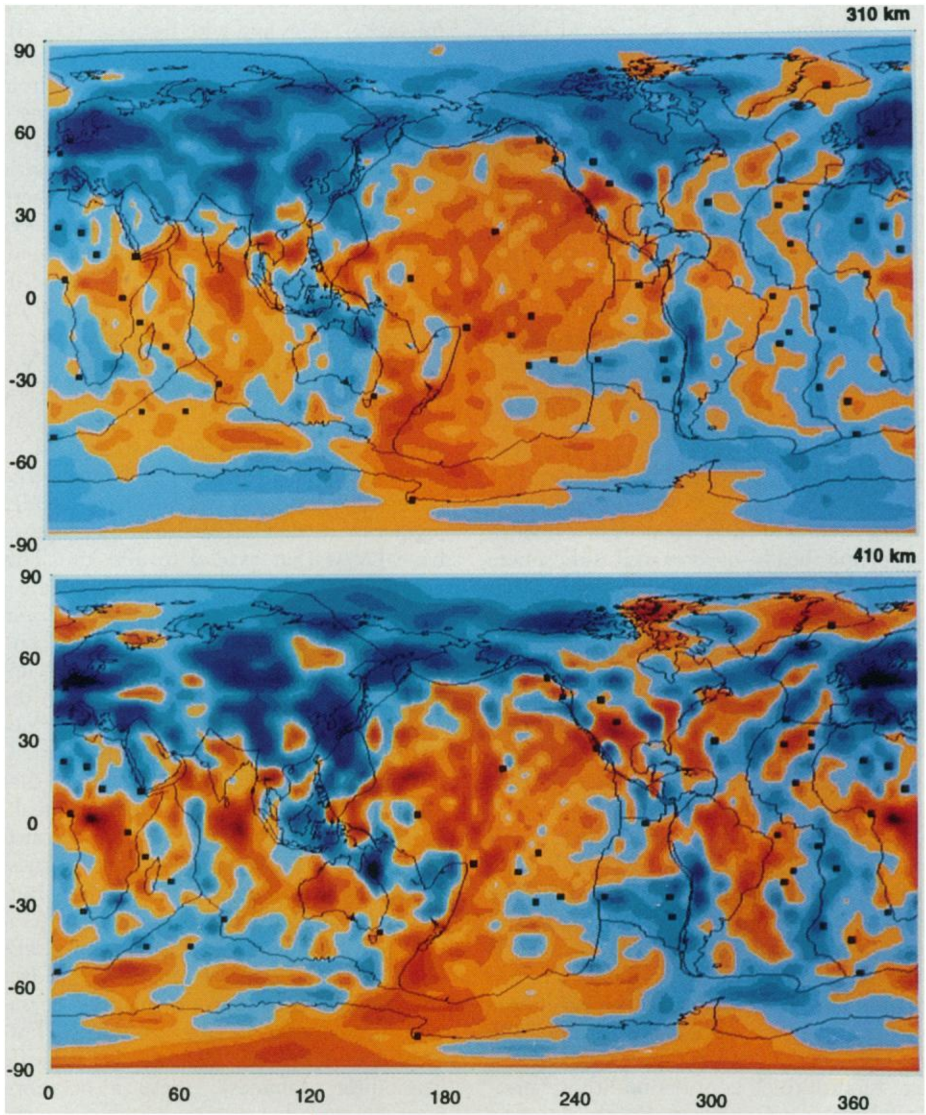

Fig. 4 (**top**). Shear velocity at a depth of 310 km. Total range is ±2.5%. Slab-cooled mantle is evident in eastern Asia, northeastern Australia, Chile, and probably central North America. Note the LVA bands in the Atlantic and Indian oceans and their association with hotspots; some of these represent former positions of migrating ridges. **Fig. 5** (**bottom**). Shear velocity at a depth of 410 km. The velocity variation is ±1%. Note the HVAs under ridges and many hotspots and the LVAs under many continental cratons. The bluest regions represent mantle cooled by current and past subduction. The Pacific Ocean mantle has not been cooled by subduction for at least 200 million years; it has not been traversed by thick cratonic lithosphere for about 500 million years. The central Pacific may not have participated in the last several cycles of continental assembly and dispersal and in the closing and opening of ocean basins.

broad, deep LVAs (Samoa, Mount Erebus, African rift) are in regions where the lithosphere is breaking, as a result of plate boundary forces, over extended regions of particularly hot mantle. These hotspots may be fed from large high-temperature upper mantle regions through lithospheric cracks, rather than from narrow, deep feeder tubes through the convecting mantle, as in the plume theories. Other hotspots over deep LVAs include Hawaii, Réunion, Canaries, Cape Verde, Kerguelen, and Tasmania. These are all nonridge hotspots. None of the on-ridge or near-ridge hotspots occur over slow (hot) mantle that extends as deep as 300 or 400 km (straight down). Hotspots in the South Pacific appear to be related to linear, shallow features of the geoid (*24*).

Continental Flood Basalts

According to recent ideas (*8, 19–21*), CFBs originate from spherical plume heads, the onset of deep mantle plumes. Others suggest that the continental lithosphere is "remobilized," "delaminated," "stretched," or otherwise damaged in the processes of CFB magmatism (*17, 25*). If these scenarios are true, there should be LVAs in the upper 100 to 200 km of the mantle under continents where CFB provinces occur, such as the North Atlantic Tertiary Province (Greenland, Scotland, Norway), Paraná-Etendeka (Brazil, southwestern Africa), Deccan (western India), Karoo (southeastern Africa), and Siberia. None of these areas show evidence for plume heads or damaged lithosphere when compared with analogous regions that have not suffered such trauma. CFBs may rather be due to lithospheric extension, or pull-apart, associated with a continent moving over hot areas (LVAs), or hot cells that may be partially molten before extension. Note the VLVAs in the North Atlantic and western Indian and southern Indian oceans where CFBs are presumed to have formed. If hotspots are initiated by plume heads, much of the asthenosphere near ridges will have enriched or hotspot characteristics rather than depleted or MORB-like (midocean ridge basalt) characteristics (*26*). Most CFB provinces occur near thick, Archean lithosphere. The large volumes of basalts in these provinces may represent ascent and melting from great depth (>150 km), triggered by extension of thick continental lithosphere.

Tristan da Cunha, a South Atlantic hotspot over which the Paraná CFB is presumed to have formed (*14, 27*), is an LVA to a depth of 210 km. Below 300 km it, and the other South Atlantic hotspots, are embedded in an extensive high-velocity, presumably cold, region. Either the associated CFBs were fed from shallow mantle or Brazil was farther to the south than current plate reconstructions suggest or the region has since been cooled by cold oceanic lithosphere sliding under South America. It is generally assumed that CFBs form over narrow fixed plumes, but there are difficulties with this interpretation (*14, 25, 27, 28*). We suggest that the Paraná CFB formed over a large South Atlantic LVA and the Gondwana flood basalt province extending across Africa, Antarctica, India, and Australia (Karoo, Fararr, Tasman) formed when this region was above the LVA in the southern Indian Ocean.

"Ghost" Ridges

There is a close association of sinuous sublithospheric LVAs with past positions of migrating ridges and with hotspots (Fig. 3). The MAR has migrated to the west since the breakup of Pangea. It decoupled from Atlantic hotspots after 70 million years ago, stranding them on the African plate (*14, 29*). At 310 km depth, the upper mantle LVA is offset to the east of the northern MAR and follows the central Atlantic hotspots. Because ridges can move more rapidly than their associated upwellings (*30*), this LVA may represent a previous location of the MAR or even the original line of continental breakup. Note also the LVA paralleling the Indian Ocean ridges. One possible implication is that spreading-induced (passive) upwelling can extend to great depth and can still feed ridges that have migrated away. Some current ridge segments in the Pacific and Indian oceans have LVAs that extend to 300 km. This may indicate long-term fixity of these ridges and a relatively deep MORB reservoir. An alternative is that continental rifts, and proto-oceanic ridges, connect active upwellings. In either case the distinction between passive-ridge and active-hotspot upwelling has become considerably blurred. Hotspots may be active upwellings that control the locations of continental breakup (*8, 20, 27*). On the other hand, continents tend to break along lines of weakness (suture zones and mobile belts). Many hotspots appear to originate under thick (cratonic) lithosphere. The associated CFBs occur near the ends of propagating ridges and the points of final continental separation, adjacent to Archean cratons. If hotspots and CFBs are due to deep mantle instabilities (plumes), these associations must be regarded as accidental. One expects ridges to propagate away from hotspots, if they are caused by active plumes, but the opposite is often observed.

The VLVAs west of the Azores and Tristan da Cunha do not contain any hotspots, perhaps because of lithospheric compression (the lithosphere is spreading, and the MAR is migrating, to the west). The Iceland and Azores LVAs are not centered under hotspots, but the islands are near the ridge, the most convenient exit. Note the sinuous LVA (Fig. 3) extending along the Atlantic hotspots (Azores–Bermuda–Cape Verde–Ascension–St. Helena–Tristan da Cunha) with branches to the Rio Grande Rise and the Scotia arc (South Atlantic), and Crozet-Kerguelen-Amsterdam (Indian Ocean). There are also LVAs paralleling the Atlantic coasts of the Americas, Greenland, and Africa. These may be due to the continental edge effect, an alternative to deep mantle plumes (*31*). The Bermuda swell and its LVA parallel the coast, which is inconsistent with the plume hypothesis. The Cameroon "hotline" also violates the plume hypothesis but is consistent with a lithospheric "crack" tapping a VLVA.

The Mesosphere

Some interesting features emerge at transition-zone depths (400 to 650 km). LVAs in the wakes of drifting continents are particularly evident for North America, India, and Antarctica, and also for South America, Australia, and Greenland (Fig. 5). The former positions of migrating ridges are also still evident, particularly the central Atlantic and the Indian ocean ridges. Upwellings lag behind if ridges migrate faster than about 1 cm/year (*30*). If ridges migrate at 10 cm/year, a typical rate, they will move 1000 km in 10 million years. Thus, there will be little correlation of ridges with deep structure, after a plate reorganization, even if ridges induce deep upwellings. Heating from below and the absence of subductive cooling are other mechanisms for generating hot upper mantle. There are LVAs, and presumably hot upwellings, in the lower mantle under the central Pacific and Africa (*3, 32*). Heat will be transferred from the lower to the upper mantle even if there is no transfer of material.

We can speculate that the linear LVA northwest of Africa may have been responsible for the Jurassic predrift volcanism along the east coast of North America (Newark group, White Mountains) and that the material from this region is now feeding the MAR and the Azores in preference to the overlying hotspots, which are on old, thick lithosphere. If the linear LVA, offset from the MAR, is a passive, or residual, upwelling, then we expect the overlying hotspots to die out as the ridge migrates farther west. The LVA under the Cameroon region (western Africa–Gulf of Guinea) may be responsible for the uplift and synchroneity of volcanism along the Cameroon "hotline." The lowest velocities are under the middle of Africa, but, at depth, the LVA broadens and underlies much of equatorial and southern Africa (Figs. 4 and 5). The low velocities may be due to continental insulation, absence of subduction, or conduction of heat from the lower mantle.

The mesosphere under Africa, Antarctica, and the Indian Ocean has low seismic velocities, and these regions are probably hot. A long-wavelength geoid high is centered over Africa, and this region may have been hot for several hundred million years (*12*). The antipodal region, the central Pacific, is also a long-wavelength geoid high and has low upper mantle seismic velocities. It is the most prolific generator of oceanic crust and oceanic plateaus on Earth. This region has not experienced slab cooling for a comparable period of time. Since the breakup of Pangea, the circum-African continents have been moving from hot mantle toward cold mantle, driven both by the upper mantle–mesospheric lateral temperature gradients and slab-plate body forces. Most of the continents have reached the end of their journey because, to proceed further, they must intrude into hot Pacific mantle. The Atlantic Ocean, and its ancestors, may never get very wide and the supercontinent cycle may not have the whole world as its stage.

The high velocities deep under Eurasia may have formed in the Triassic when microcontinents were assembling and ocean basins were closing. The very low seismic velocities and presumably low viscosity under the Indian-Australian plates may explain why this region is moving rapidly. Superposed on the thermal currents in the sublithospheric mantle are plate-induced motions, generally directed from ridges to trenches. Lateral temperature gradients may have driven continental fragments north from Gondwana to Laurasis, across the Tethys Sea (*33*).

Cratons

Most continents have Precambrian nuclei that have been stable for more than 2 billion years. The continental lithosphere under this ancient crust is cold and thick (Figs. 1 through 3). The largest or most pronounced seismically defined cratons are the Canadian, Baltic-Ukranian, and Western Australian shields. Others are evident in Africa, Brazil, India, Greenland, and Antarctica. Some of the fastest areas are under ice caps or sediments, suggesting that the oldest crust may yet to be found. Figure 2 shows velocities at 110 km, which is below the base of the elastic lithosphere in all parts of the world except cratons. The fastest regions, without exception, are associated with Archean shields. The Indosinia Block (Cambodia) and Arabia are the only shields that are not HVAs. Even the small Tarim shield (north of India) shows up as an HVA. This appears to be a strong, rigid block. It appears that ancient continental lithosphere may extend to about 200 km. The seismic high-velocity layer and the flexural elastic thickness of the lithosphere may approximate the thickness of the plate (*13, 34*). Many noncratonic areas (eastern and southeastern Asia, northern Australia, Arctic Ocean, South Atlantic) have fast velocities and some cratons (South America, India, Arabia, central Africa) have relatively low velocities below about 220 km. Some cratons have overridden cold oceanic lithosphere, and this can be confused with a continental root.

Eurasia is a landmass created by the closure of small oceans and the amalgamation of about eight minicontinents, some with cratons. It has not moved far since its assembly as the northern part of the supercontinent of Pangea, and it is therefore still underlain by foundered oceanic lithosphere. The high seismic velocities found throughout the upper mantle under Eurasia (Fig. 5) suggest that the mantle here has been cooled by subducted slabs. These slabs may still reside in the mesosphere.

Tectonically Active Areas: Lithospheric Extension Versus Deep Plumes

Earthquake-prone and volcano-prone continental areas are embedded in vast low-velocity, high-temperature domains (Figs. 1 and 2). Western North America is on the edge of a VLVA that extends across the Pacific Ocean. Other boundaries of this system include eastern and southeastern Asia, eastern Australia, and New Zealand, all tectonically unstable regions. Other active regions (southern Europe, northeastern Africa, western Antarctica) also occur at the edges of VLVAs. The high inferred temperatures under southwestern United States may have caused tectonic and magmatic activity as North America drifted over this region. Uplift and volcanism are sometimes attributed to the arrival of a giant plume head from the core-mantle boundary (*8, 19, 20*) but could be the result of drift and extension over preexisting hot mantle.

The Rhinegraben and Lake Baikal rifts exhibit only moderate volcanism. They have low seismic velocities above 200 km. Below 300 km, both regions appear to be cold. The same is true, to a lesser extent, for most of the Red Sea. These may or may not be hotspots (*27, 35, 36*). It is likely that the Rhinegraben and Lake Baikal are in regions of externally induced tension, possibly due to collision of Africa and India, respectively, and that volcanism is entirely passive rather than related to plumes or sublithospheric LVAs. The Rhinegraben Eiffel "hotspot" does not have a time-progressive track (*37*). Other rifts (eastern Africa, Gulf of California, Gulf of Aden, Ross Sea) are over deep hot mantle. These are in diverging regions rather than in regions of collision-induced rifting. The question then is, does rifting cause the upwellings and massive volcanism we call "hotspots," or do deep mantle plumes cause rifting and hotspots? The association of hotspots with shallow LVAs, Pangea, spreading centers (past and present), continental wakes, lithospheric extension, and absence of slabs is fairly impressive. Not all rifts or hotspots have deep LVAs. These observations are all consistent with lithospheric control on the locations of some, if not most, hotspots and a general control by post-Pangeatic subduction on high velocities and the absence of hotspots. "Hot" mantle may be "normal" mantle, uncooled by subduction.

Many deep LVAs are unrelated to hotspots. The lowest velocity regions are beneath southwestern North America, southern Marianna Arc, New Zealand, southeastern Asia, Indochina, northern Indian Ocean, and Gulf of Aden. The lithosphere in many of the above areas is in tension. The Afar appears to lie at the edge of an enormous hot area. The Saudi shield is the northern boundary of this VLVA, which underlies much of Africa and the Indian Ocean. Generally, regions of subduction have fast seismic velocities below about 300 km (*2, 38, 39*) (Fig. 5). The low velocities under New Zealand–Tonga–Fiji and Mariannas are therefore surprising. The LVAs could have resulted from recent arc-rise collision, backarc spreading, or slab dehydration. There is also no depression of the 650-km discontinuity here, as one might expect under a cold slab (*40*).

Convergence and Subduction: Past and Present

Oceanic plates cool at the surface of Earth before plunging back into the interior. Because they are cold and have a long thermal time constant, they should show up as HVAs even after residing in the mantle for hundreds of millions of years. HVAs are prominent (Figs. 3 to 5) where subduction has been long-lived (Kamchatka, Japan, Borneo, Sumatra, northern Australia, New Hebrides). Also note the Alpine belt, the Andes, the Bering Sea, and Asia north of India. A belt of HVAs extends from Canada through North America to South America (Figs. 4 and 5), where the active margins of the Americas were some 60 to 80 million years ago. Before this time, North America was overriding thick oceanic lithosphere. Subsequently, flat subduction of hotter lithosphere was responsible for the Laramide orogeny. The HVA from Italy to Arabia (Fig. 5) is possibly due to closure of the Tethys Ocean.

The mesosphere may be the region where slabs accumulate (*13, 41–43*). The boundaries of the mesosphere (400 and 650 km) represent changes in mineralogy, but they need not be equilibrium phase changes in a homogeneous mantle. There is little correlation of seismic velocities (Fig. 5) with present surface tectonics but good correlation with post-Pangeatic subduction (*12, 13*). Subduction zones generally mark the lateral boundaries between slow and fast mantle, for example, Tonga-Fiji, Mariannas, Peru, Cascadia, and the Aleutians (Fig. 5). Below a depth of 300 km, the fast regions generally correlate with expected locations of oceanic lithosphere overridden by continents since the breakup of Pangea (*12, 13*). The HVA in northeastern Australia is probably due to subduction of the northern New Guinea plate. The high velocities in eastern, southeastern, and southern Asia are the result of subduction over the past 230 million years. The high velocities under Eurasia may be due to even more ancient subduction.

Discussion

Tomography allows us to look beneath the lithosphere and back in time. The mantle has a long memory because of the long conductive time constant of subducted lithosphere and the persistence of convective upwellings and cell boundaries. Subduction, an instability in the surface boundary layer, induces downwellings in the underlying mantle. Old subducted plate cools off the surrounding mantle, generating high-velocity anomalies, sluggish convection, and cold cells. Mantle flows toward regions where plates spread apart; upwelling and melting at these places are caused by plate processes, not thermal instabilities. Directions of plate motions change as a result of changes in boundary conditions (collisions of continents, ridges, arcs, and oceanic plateaus), and this causes new rifts to open and others to close. Midocean ridges respond by migrating. Previous locations of passive upwellings are evident in the tomography, as are old positions of convergence and subduction. The plates appear to control the long-wavelength tomography and the planform of mantle convection. Many hotspots are related to spreading ridges, both present and past, raising the question of whether they are caused by passive (rifting) or active (thermal instability) processes. There are many hot regions in the mantle that are unrelated to hotspots. Some hot regions appear to be hotter than average convection cells, because of their prior history, rather than deep mantle plumes.

It is generally assumed, perhaps erroneously, that geochemically depleted (*26*) basalts come from the shallowest mantle even though enriched basalts erupt at most times and places where one expects this mantle to be sampled; the most depleted basalts are associated with long-lived and rapid-spreading conditions that favor a broader, deeper source volume. Enriched basalts dominate at the initial stages of lithospheric disruption and are a common component in most basalts. These are usually attributed to deep mantle plumes that pierce the depleted asthenosphere. We favor an inhomogeneous upper mantle that is contaminated by subduction and trapped melts and flushed out by volcanism. The main basaltic reservoir is probably below this contaminated layer, but material becomes contaminated upon ascent, particularly away from mature ridges. Sustained upwellings can push the enriched layer aside. It will be of interest to compare the geochemical characteristics of ocean ridge and hotspot magmas with the tomographic patterns discussed here.

Stable Archean plates appear to be about 200 km thick. Some have overridden cold oceanic lithosphere or coalesced with others by the elimination of intervening oceanic plates. These regions are cold throughout the upper mantle. Cold and hot convection cells are alternatives to deep continental roots and deep mantle plumes as explanations for lateral temperature variations. Individual convection cells may be well mixed, but it is unlikely that all convection cells in the mantle have the same history, composition, and temperature. Plume theories focus on a more-or-less isothermal homogeneous mantle. The tomography exposes pronounced intrinsic lateral variations.

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The Tropical Timber Trade and Sustainable Development

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The tropical timber trade appears to have promoted neither sustained forest management nor sustained forest-based industrialization. The boom-and-bust export pattern is often blamed on demand by developed countries, high import barriers, and low international wood prices. In fact, it is rooted in tropical countries' own policies related to timber concessions and wood-processing industries. These policies suppress timber scarcity signals and must be revised if the trade is to promote sustained economic growth. Even if this is done, the trade may not promote sustained-yield forestry in individual countries.

The history of the tropical timber trade is discouraging both to foresters and environmentalists interested in sustained management of tropical forests and to policy-makers interested in sustained industrialization in the forest sector. Since the end of World War II, one tropical country after another has followed a boom-and-bust export pattern (*1–3*): High initial export earnings are followed by depletion of old-growth forests, a lack of management of second-growth forests (*4*), and a collapse of domestic processing industries. Logging and processing industries enjoy profits during the boom, but the economic activity is not sustained.

This pattern emerged in West Africa in the 1950s and 1960s. It became even more apparent in the 1970s and 1980s as the trade shifted toward Southeast Asia and expanded in volume. In Southeast Asia today, several countries have already gone bust (for example, Thailand and the Philippines), others will shortly (for example, the state of Sabah in Malaysia), and in most remaining countries the boom is either cresting or waning (*1, 5*).

Is the boom-and-bust pattern inevitable? If so, is the tropical timber trade inherently incompatible with sustainable development? International timber prices reflect the commercial value of tropical wood—not the diverse values of tropical forests as sources of biological diversity, clean water, and nontimber forest produce. Nevertheless, can the trade indirectly protect these nonmarket values by generating incentives to maintain permanent forest areas?

This article provides an economic perspective on these issues (*6*). Although the timber trade provides opportunities for a tropical country to enhance its overall economic performance, the trade does not necessarily create incentives for sustained forest management or for sustained industrialization within the forest sector. Policies in tropical countries have generally reduced the economic benefits that those countries can reap from the trade and have reduced the likelihood that the trade can promote sustainable development of the forest sector.

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Misconceptions About the Trade

The inability of tropical timber-exporting countries to break out of the boom-and-bust pattern is often attributed to three factors: developed countries' exploitation of tropical countries' timber resources (*7, 8*), high import barriers by developed countries against processed tropical timber products (*9*), and low prices for tropical timber in international markets (*7, 8*). Consumption in developed countries allegedly drives the boom. Import barriers allegedly inhibit the development of processing industries in tropical countries, reducing those countries' export earnings and the value of their forests as a source of raw materials. Low prices allegedly reflect market manipulation by developed countries and reduce the financial viability of forest management.

None of these three factors holds up well when trade statistics are examined (Table 1). In 1989, developing countries (excluding China), which are mainly tropical, exported 11% of their harvest of industrial roundwood. They exported 23% of their output of solid-wood processed products, and smaller percentages of their output of fiber products. Taken together, these figures indicate that only about a third of the industrial roundwood harvested in developing countries entered international trade in any form. Moreover, many of the exports were to other developing countries (*10*).

In 1989, developing countries (excluding China) imported only a slightly smaller value of wood products than they exported, \$11.5 billion versus \$12.7 billion (*10*). There is a significant international flow of tropical solid-wood products, which are mainly hardwood (nonconiferous), from developing to developed countries, but