Observational hints for a plume-fed, suboceanic asthenosphere and its role in mantle convection

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Abstract. An asthenosphere layer which is entirely fed from below by plumes and which loses equal mass by accretion to the overlying oceanic lithosphere and at subduction zones may play a critical role in shaping the form of mantle convection. In this study we discuss geochemical, seismic, and geoid(depth) evidence for lateral flow within this type of asthenosphere. In particular, we suggest that there are large-scale layered, horizontal flow structures that connect upward plume input beneath hotspots to near-ridge regions of increased asthenosphere accretion into the growing oceanic lithosphere. Lateral asthenosphere flow is also shaped by oceanic subduction zones, with a partial return flow from trenches, and by deep continental roots that are migrating barriers to asthenosphere flow. This alternative paradigm offers relatively simple explanations for several puzzles about mantle convection, for example, the low mantle heat flow beneath continents. It also offers an explanation for why mid-ocean ridges appear to be passive features that migrate with little geochemical or morphological change with respect to the lower mantle and seem to be uncoupled from large-scale mantle flow, while in contrast, trenches appear to be strongly coupled to mantle-thick regions of fast (colder) seismic velocity anomalies. We also discuss several implications of this paradigm that should be testable in future studies, such as the prediction of cogenetic off-axis seamount volcanism that is created between an off-axis hotspot and its neighboring ridge axis.

Introduction

At the dawn of the plate tectonics era came the realization that oceanic lithosphere forms the cold top thermal boundary layer of the convecting mantle [Elsasser, 1969; Morgan, 1968; Richter, 1973]. Early cartoons illustrating this process typically showed closed convection cells of upwelling mantle beneath mid-ocean spreading centers and downwelling lithosphere and mantle at oceanic subduction zones (Figure 1a). These cartoons may arise because this cellular convection pattern is a typical feature of constant viscosity numerical simulations of mantle convection [e.g., McKenzie et al., 1974]. Until recently, only these types of simple viscosity structures could be explored by numerical experiments. Later, we will discuss recent work exploring more complex temperature- and stress-dependent mantle rheologies. However, there are several fundamental problems with this scenario. The surface relief predicted by this type of convection model [Davies, 1988] is quite different from the average seafloor depth-depended that is both observed [Parsons and Sclater, 1977; Smith, 1990] and predicted by models of a simple cooling thermal boundary layer [Langseth et al., 1966; Vogt and Ostenso, 1967; Davis and Lister, 1974]. This scenario also predicts that large, 150°-300°C [Davies, 1988] lateral temperature variations will exist in the upper mantle. Mid-ocean ridges migrate over the upper mantle, typically with velocities of a few centimeters per year. If mid-ocean ridges migrate with respect to mantle convection cells, this would lead to large variations in oceanic crustal thickness as ridges migrate over these hot and cold anomalies. The predicted variations in crustal thickness are larger than those observed. Except at on-ridge or near-ridge hotspots like the Iceland hotspot, seismic observations of oceanic crustal thickness appear to cluster close to 6-7(±2) km [Chen, 1992; White et al., 1992]. This variation in crustal thickness from ~4-9 km is likely to be close to an upper bound on crustal thickness variation, because both the thickest and thinnest crustal measurements occur at slow-spreading ridges [Chen, 1992]. Local crustal accretion at slow-spreading ridges is strongly influenced by the segmentation pattern of the ridge axis, with thin ~4-km thick crust found beneath transform and nontransform offsets, and thicker crust seen near the shallow center of the ridge segment [Detrick et al., 1993; Tolstoy et al., 1993; Lin and Phipps Morgan, 1992]. Thus the segment averaged crustal thickness inferred from regional gravity surveys [Blackman and Forsyth, 1991] is likely to be much closer to 6-7(±0.5-1) km than is implied by the ±2-km estimates of the scatter in "point" seismic measurements of seafloor created at slow-spreading ridges. Even ±2-km variations in oceanic crustal thickness are considerably smaller than those that would be produced by 150°-300°C variations in upwelling mantle beneath a spreading center. For
Figure 1. (a) Pattern of mantle flow seen in many convection cartoons where surface plate motions are part of a single mantle flow circulation. In this case plate and mantle flow are tightly coupled beneath suboceanic and subcontinental lithosphere, which leads to several geological implications which are not observed, as discussed in the text. (b) Pattern of mantle flow if a suboceanic asthenosphere layer acts to decouple surface plate motions from deeper mantle flow. In this case plate and mantle flow will only be coupled at oceanic subduction zones where the lithosphere penetrates beneath the asthenosphere layer, and perhaps beneath thick continental cratons where no asthenosphere layer is present. We propose that the weak asthenosphere layer is preferentially fed by upwelling plumes from the lower mantle. This leads to concentrated pipes of flow from plumes to the neighboring ridge axis, shown by the dotted arrows. Elsewhere, asthenosphere flows to replace material which is accreted to the overriding lithosphere. The circuit of mantle transport is completed by lithosphere subduction to the deeper mantle. See text for further discussion. (c) The influence of the plume upwelling and asthenosphere decoupling of lithosphere motions from deeper mantle flow could be even more profound. The cartoon shown here shows a scenario where the net mantle motion is downward except in areas of plume upwelling. The mantle temperature is controlled by the "ice cube" effect of subducted slabs, radiogenic mantle heating, and heat input from the core across the core-mantle boundary (CMB). Even though the mantle is the source for most of the heat lost at the Earth's surface, plume flow could act to "short-circuit" the convection process so that most mantle heat is transported first down to the CMB and then up plume pipes to feed the asthenosphere. This scenario is meant to provoke. Its feasibility will be readily testable once we can incorporate plumes and the rheological structures of lithosphere, asthenosphere, and D" boundary layers into a three-dimensional convection code.
example, mantle peridotite in the garnet lherzolite stability field has a pressure dependence to its solidus temperature of $2^\circ$-$3^\circ$C/km (Hess, 1992) and a latent heat of fusion that, in terms of the temperature "reduction" from melting, is $\Delta h_{f} = 500^\circ$-$600^\circ$C$\cdot\Delta f$, where $\Delta f$ is the increment of melting. An excess temperature anomaly of $150^\circ$C would correspond to melting starting $50$-$75$ km deeper beneath the ridge. This deeper melting would then translate into an additional (deep) increment of melting $\Delta f = 50$-$500$ km/s $\Delta f = 0.25$-$0.30$ during ascent beneath the ridge. Most petrological and geological studies suggest that the total extent of melting $f$ beneath a ridge is less than $f = 0.15$-$0.3$ (Klein and Langmuir, 1987; Hess, 1992; Langmuir et al., 1992; Forsyth, 1993). A variation in $f$ of $0.25$-$0.30$ above a minimum melting fraction of $0.15$ would correspond to variations in crustal thickness by a factor of $0.40/0.15$-$0.45/0.15 = 2.6$-$3$. This predicted variation is at least $\sim 50\%$ higher than that observed. Instead, the observed variation of oceanic crustal thicknesses suggests that the lateral upper mantle temperature variations sampled by the mid-ocean system are likely to be less than $\sim 75^\circ$-$100^\circ$C, unless colder mantle temperatures are tightly linked to compositionally fertile (easy to melt) mantle compositions.

**Evidence for a Weak Asthenosphere Layer**

The "standard model" in Figure 1a is based on the assumption of a uniform viscosity mantle, or a $\sim 700$-km-thick upper mantle that is $\sim 30$ times less viscous than underlying lower mantle. However, several geophysical observations seem to imply the existence of a low-viscosity decoupling zone between the oceanic lithosphere and deeper mantle. Seismic observations of radial Earth structure have long supported the existence of a low-velocity [Dziewonski and Anderson, 1981] and high attenuation [Widmer et al., 1991] "asthenosphere" layer between $\sim 100$-$300$ km depth. Complementary seismic evidence of a relative high-velocity shallow mantle under continents [Jordan, 1981; Su et al., 1992; Zhang and Taninomo, 1992] argues that this asthenosphere layer is largely confined beneath oceanic lithosphere. Rock mechanic arguments suggest a low-viscosity zone at the shallowest hot mantle where pressure effects lead to a viscosity minimum along a mantle adiabat [Weertman and Weertman, 1975; Buck and Parmentier, 1986; Karato and Wu, 1993] (Figure 2). Studies of the distribution of stresses in the oceanic plates support this; they require a low-viscosity asthenosphere of $10^{18}$-$10^{19}$ Pa s underneath the oceanic lithosphere [Richter and McKenzie, 1978; Wiens and Stein, 1985]. This is significantly lower than the values of $10^{21}$ Pa s inferred for the upper mantle beneath continents from investigations of glacial rebound [Peltier, 1989], but consistent with the $10^{18}$-$10^{19}$ Pa s viscosities inferred from a glacial rebound experiment in Iceland [Sigmundsson and Einarsdottir, 1992]. Geomagnetic inversion studies also favor an enhanced electrical conductivity at this $\sim 100$-$300$-km depth [Oldenburg, 1981; Constable, 1992].

How weak does the asthenosphere need to be to significantly modulate mantle flow? For horizontal flow in a channel of thickness $h$ and fluid viscosity $\eta$, the relation between the net horizontal flux $\dot{q}$ and the lateral pressure gradient $\nabla p$ is $\dot{q} = h^2/12n |\nabla p|$. This relation implies that horizontal flow in response to a lateral pressure gradient will preferentially concentrate into a $\sim 200$-km thick mantle layer instead of the entire $\sim 600$-km thick upper mantle (less lithosphere) if the asthenosphere layer is $\sim 30$ times less viscous than the upper mantle as a whole. This is likely to be the case. Olivine is both the most abundant and the weakest major mineral under a wide range of upper mantle conditions; thus the rheology of the uppermost mantle is probably close to that of polycrystalline olivine. At steady state the rheology of polycrystalline olivine depends on temperature, pressure, grain size, and shear stress. Karato and Wu's [1993] recent summary of the rheology of the uppermost mantle is that for the most probable viscous deformation mecha-
nism in the uppermost mantle, dislocation creep of olivine, the viscosity \( \eta = \eta_0 e^{\frac{(E^* + PV^*)}{RT}} \), where \( \eta \) is the shear stress, \( m \) is the grain-size exponent, \( \alpha \) is the stress exponent, \( \beta \) for dislocation creep, \( E^* \approx 540 \text{ kJ/mol} \), \( V^* \approx 6 \times 10^{-6} \text{ m}^3/\text{mol} \) is the activation volume, and \( R \) is the ideal gas constant. In this case, increasing pressure alone would lead to a 200-7000-fold increase in upper mantle viscosity between 150-km and 300-km depth. Karato and Wu [1993] suggest that for viscosities larger than \( 10^{21} \text{ Poise} \), diffusion creep (with smaller pressure dependence \( V^* = 6 \times 10^{-6} \text{ m}^3/\text{mol} \)) will become the dominant creep mechanism, thus limiting the maximum viscosity increase through the asthenosphere to \( 1000 \)-fold instead of \( 7000 \)-fold. The lower activation volume \( V^* = 10^{-6} \text{ m}^3/\text{mol} \) that is used by Buck and Parmentier [1986] still leads to a pressure-induced viscosity increase of 35-fold for a depth increase of 150 km. If the asthenosphere is actually hotter than underlying mantle (Figure 2, see below), then this temperature difference will increase the viscosity contrast between asthenosphere and underlying mantle. Even for a low activation volume of \( V^* = 10^{-6} \text{ m}^3/\text{mol} \), a 200°C decrease in temperature between a hotter plume-fed asthenosphere at 150-km-depths and "normal mantle" at 300-km-depths would imply 4 orders of magnitude reduction in asthenosphere viscosity due to the combined effects of elevated asthenosphere temperature and reduced confining pressure with respect to underlying mantle.

Plume-Fed Asthenosphere?

The original plume theory [Morgan, 1971; 1972] suggested that the asthenosphere may be preferentially fed by hot material which lies on a direct adiabat from a core-mantle boundary layer (Figure 2), leading to a hotter and lighter layer which stably overlies colder deeper mantle. This idea, that \( 25-40 \) plumes are the sites where deep upwelling is strongly concentrated [Morgan, 1971], has been largely neglected in mantle dynamics because it appears to be contradicted by heat flow observations. A simple heat flow argument suggests that plumes cannot be the source of much upwelling mantle because hotspots directly account for only 5-10% [Davies, 1988; Sleep, 1990] of the surface heat flow, much less than the \( \approx 80\% \) of the Earth's interior heat which is lost through lithosphere accretion away from mid-ocean ridges. This argument may be wrong. It would be correct only if hotspots (the volcanic surface expression of an underlying plume) and ridges were fed by different sources of mantle upwelling. If, instead, upwelling mantle from plumes flows laterally in the asthenosphere to feed mid-ocean ridges, then plumes could be the source of all upwelling and upward mantle heat transport, heat which is lost to a small degree at hotspots and to a much larger degree by lithosphere accretion (and plate subduction and the thermal reincorporation of subducted lithosphere).

Previous Work on the Role of the Asthenosphere in Mantle Convection

Thinking about an asthenosphere layer dates to the beginning of plate tectonics. A passive weak asthenosphere layer was proposed to somewhat decouple lithospheric and lower mantle motions [Richter and McKenzie, 1978; Wiens and Stein, 1985]. Dynamic flow within a global asthenosphere layer that channels the return flow from subduction zones to ridges was explored in several studies in the 1970s [Schubert and Turcotte, 1972; Schubert et al., 1976; 1978; Yuen et al., 1978; Parmentier and Oliver, 1979; Chase, 1979]. These studies have been largely neglected because this counterflow appears to predict that old seafloor near subduction zones should be shallower than young seafloor near ridges [Schubert and Turcotte, 1972], in gross contradiction with observations that the seafloor generally deepens as it ages and cools away from a ridge crest.

Recently investigators have begun to use numerical experiments to explore the effects of three-dimensional mantle flow [Cseres and Christensen, 1990; Christensen and Harder, 1991; Rabinowicz et al., 1990; 1993; Honda et al., 1993; Tackley et al., 1993], and to explore the effects of temperature- and stress-dependent viscosity laws [Christensen and Yuen, 1989; Malesic et al., 1992; Christensen and Harder, 1991] on the structure of convection. These recent numerical studies support observational evidence that a shallow asthenosphere layer is not the dominant region for return flow from subduction zones to spreading ridges; instead, the deeper mantle must be included in the return flow to feed plate accretion [Rabinowicz et al., 1990]. A recent study has found that a 1000-fold drop in asthenosphere viscosity relative to underlying mantle will effectively decouple asthenosphere flow from underlying convection [Rabinowicz et al., 1993]. However, neither this recent study nor any other study to date has explored the effects of flow within a model that has a plume-fed, weak asthenosphere layer as the geometry for return flow from the deeper mantle to feed lithosphere accretion. This remains a formidable computational task that, at last, lies near the current state-of-the-art for numerical experimentation.

Proposed Asthenosphere Structure and Role

Here we explore a variant of these earlier proposals: in our model the oceanic asthenosphere layer is entirely fed by plumes from the deeper mantle and consumed entirely by cooling and underplating onto the overlying lithosphere (the most rapid consumption occurs beneath young seafloor near rise crests), with subsequent subduction of the lithosphere to the deeper mantle. Beneath continents a deep continental "root" may preclude the existence of any asthenosphere layer. In this scenario the asthenosphere plays a different role in modulating mantle convection: it acts as a spatially discontinuous weak zone between plate motions and deeper mantle flow (which couple directly at subduction zones and possibly the deep roots to continental cratons), and it acts as a region of large lateral flow from hotspots [Morgan, 1971] (and possibly, partial return flow of asthenosphere from subduction zones [Phipps Morgan and Smith, 1992]) toward ridges to supply the material needed for lithosphere accretion. After presenting observational evidence that this type of asthenosphere may actually exist in Earth, we will discuss several potential implications of this form of asthenosphere flow on volcanism, seafloor depth, and geoid anomalies, and on mantle convection.

Geochemical Evidence for Lateral Plume-Ridge Flow

There is strong geological and geochemical evidence that there is a flow connection between off-axis hotspots and neighboring mid-ocean ridges. Volcanism and island building appear to be enhanced along a mid-ocean ridge at the places that are closest to a hotspot, this phenomena was described by [Morgan, 1978, p. 3355] as a "second type of hotspot island", and attributed to a pipeline channeled flow from the hotspot that met the ridge below this region of enhanced volcanism. Schilling and coworkers noted in a series of pioneering geochemical papers [Schilling et al., 1979; Schilling, 1986; Humphris et al., 1985; Hanan et al.,]
that there is a geochemical "spike" of a more hotspotlike isotopic signature on the ridge segment that is closest to the hotspot and proposed that this spike was due to a pipe flow from the hotspot to the nearest ridge. This signature is clearly seen in the South Atlantic [Humphris et al., 1985; Hanan et al., 1986], where many off-axis hotspots appear to have an associated on-ridge isotopic spike. It is also seen in isotopic and trace element ratios in the North Atlantic. Here along-axis "highs" in isotope and trace element ratios are seen in Figure 3a [Dosso et al., 1993] to be associated with the on-ridge Iceland hotspot, the near-ridge Azores hotspot, and the far-from-ridge Cape Verde (and Canary?) hotspots. Additional intriguing evidence has been recently seen along the East Pacific Rise near 17ºS [Mahoney et al., 1994], where a clear isotopic spike in the mid-ocean ridge basalts (MORB) is seen. Figure 3b shows the He and Sr isotope data; the lead and Nd isotope systematics parallel those of Sr. (Curiously, here trace element observations do not apparently correlate with isotope systematics [Mahoney et al., 1994].) We suggest that this spike may reflect eastward lateral flow to the East Pacific Rise from the Society hotspot province that lies several thousand kilometers to the west [Mahoney et al., 1994]. This possibility has intriguing support from seismic evidence that we discuss below.

**Asthenosphere Flow Patterns and the Structure of Geochemical Anomalies**

The distribution of geochemical heterogeneities within the upper mantle is somewhat uncertain, with strong evidence for significant trace element and isotopic heterogeneity on length scales from ~1-10-m scales seen in lherzolite outcrops of presumed upper mantle rocks [Pole and Allegre, 1981] to the ~10,000-km-scale Dupal anomaly mapped in the geochemical distributions of oceanic basalts [Zindler and Hart, 1986].

We prefer the working hypothesis that most of the trace element and isotopic variation in hotspot or ocean island basalts (OIB) and MORB is due to the incorporation within basaltic magmas of different amounts of material created by prior differentiation processes (e.g. prior OIB and MORB melting, sediments, all somewhat modified by partial melting at a subduction zone during "recycling" into the lower mantle reservoir for upwelling plumes [White and Hofmann, 1982; Hofmann and White, 1982; Davies, 1984; Allegre and Turcotte, 1986]). These concentrated plums of trace element and isotopic heterogeneity will be preferentially sampled and removed by plume melting near the hotspot because they have lower melting temperatures than the bulk of the harzburgite/lherzolite material which is brought up at.

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**Figure 3a.** (a) Along-axis bathymetric [Vogt, 1986] and geochemical [Dosso et al., 1993] variation along the northern Mid-Atlantic Ridge. Note that almost all geochemical "spikes" along the ridge are associated with ridge segments that are closest to a neighboring hotspot. This interpretation holds for the on-ridge Iceland hotspot, the near-ridge Azores hotspot, and the far-from ridge Cape Verde hotspot.
a plume. During the phase of lateral asthenosphere transport in a pipe from the hotspot to the ridge, little ascent and pressure release melting will occur. Thus the initial upwelling and melting near the ridge would continue to melt-out the plums of heterogeneity that survived the hotspot melting process resulting in a geochemical "spike", similar to that of the associated off-axis hotspot, where an asthenosphere plume-to-ridge pipe intersects the spreading system.

A spreading center is likely to be a further region of concentrated along-axis asthenosphere flow, because of the fluidity of subridge asthenosphere relative to asthenosphere beneath older seafloor. The temperature and pressure dependence of mantle viscosity leads to smallest viscosities where asthenosphere is both hot and shallow, as it is directly beneath the thin thermal boundary layer at a ridge. In addition, the presence of partial melt within and above the region of ridge melting may further weaken the asthenosphere by a factor ranging from $-3$ [Kohlstedt, 1992] to $-10$-$50$ (G. Hirth, personal communication, 1993). The spreading center channel for asthenosphere flow differs from a hotspot-ridge pipe in that flow is preferentially channeled at shallower depths where low-melting-point geochemical heterogeneities are likely to have been removed by prior pressure-release melting. Further melting during the concentrated along-axis flow of asthenosphere within the weakest shallow levels is likely to melt material that is relatively depleted, thus pooled melts collected above a given section of the ridge may have a major element chemistry that is dominated by melting of depleted asthenosphere that has flowed along-axis into its site of melting while the trace element and isotope signature of MORBs derived by pooled melts may come from a deeper, more viscous and more slowly moving asthenosphere source. The fact that pressure release melting beneath a given section of the ridge only samples the vertical velocity field within the region of melting, while the melting history may depend on the path of lateral flow introduces a complexity to the major and trace element geochemical history that cannot be studied using the residual melting column approach developed by Klein and Langmuir [1987], Langmuir et al. [1992], and Plank and Langmuir [1992].

In our proposed scenario the final phase of asthenosphere flow from the spreading center to where asthenosphere is consumed by off-axis plate accretion is the only phase of flow where the asthenosphere generally moves in the direction of the overriding plate. Nonetheless, this flow pattern may be the most prevalent spatial pattern simply because the flow in plume-to-ridge pipes and subsequent along-ridge flow is concentrated into localized regions. If this flow occurs beneath progressively thickening lithosphere, then it is likely that this phase of flow will take place with the least opportunity for ascent and pressure release melting.

Seismic Evidence for a Suboceanic, Plume-Fed Asthenosphere

Seismic surface wave velocity inversions provide additional hints that this type of plume-fed asthenosphere is likely to exist. Next we will discuss the implications of several vertical cross sections shown in Figure 4 that are extracted from the recent 5°x5° upper mantle shear velocity model RG5.5 [Zhang and Tanimoto, 1993]. These cross sections have a horizontal resolution of $-1000$ km. Zhang and Tanimoto [1992] showed the existence of many pipelike connections between Atlantic plumes and the Mid-Atlantic Ridge, as illustrated by the example in Figure 4. Even more striking are hints of connections between major Pacific plumes and the East Pacific Rise (EPR). Consider the Tahiti hotspot-EPR or Hawaii hotspot-EPR cross sections shown in Figure 4. In each case we see the deepest thermal signature beneath the hotspot and a hot, slow zone connecting the plume and ridge. Note in particular that the thermal anomaly is not rafted passively by plate shear which would move it to the northwest, but instead appears to be moving east-southeast towards the ridge.
In our preferred scenario for asthenosphere flow, in general, asthenosphere flow should move from a plume source towards a ridge because most asthenosphere is consumed by accretion to the growing lithosphere near a mid-ocean ridge where age plate thickening is most rapid. If melting near a spreading center results in the creation of a thick, strong compositional layer above the depth of the melting region, then near-ridge asthenosphere accretion into lithosphere will be even stronger than age plate cooling. We can also see in Figure 4 the absence of a plume root for a Pacific profile D-D' which does not cross any known hotspot; in this case hot asthenosphere is more restricted to a broad region about the EPR. (There are small anomalies where the profile crosses fossil hotspot swells.)

Other general structural features of these cross sections are the strong horizontal layering of seismic velocity anomalies (readily apparent despite the strong vertical exaggeration in each cross section) and the shallowing of the slow asthenosphere region near mid-ocean ridges. This latter feature could be a consequence of the asthenosphere flow being mostly confined to the weakest rheological channel which should lie just below the oceanic lithosphere. Profile B shows a reduction in the magnitude of the slow hotspot anomaly between the slowest shear ve-
velocity structures beneath Hawaii and the near the ridge. This reduction may be real, but may also reflect variations in the path density and coverage of the underlying surface wave data. It is intriguing because this type of reduction would be anticipated if the largest slow anomalies are due to the presence of partial melt. Since pressure release melting is limited to regions of upwelling, melting near the hotspot would be much larger than subsequent melting during predominantly lateral asthenosphere flow towards the ridge, with another increase in melting during upwelling near the ridge. During the period of lateral ridgeward flow, gradual escape of a residual melt fraction created near the hotspot could lead to an increase in seismic velocity. This scenario is consistent with the velocity structure seen in profile B of Figure 4.

Since the inferences above depend on the ability of high-resolution surface wave inversions to image relatively short-wavelength (~1000-2000 km) structures, the question of resolution is key to our belief in the conclusions drawn from Figure 4. Developing better seismic images of upper mantle shear velocity structure is a rapidly progressing field, and we are aware of two (as yet unpublished) efforts by Ekstrom and Tromp (G. Ekstrom, personal communication, 1993) and Laske and Masters (G. Laske, personal communication, 1994) that will help to better assess these questions. (It is encouraging that both of these latter inversions appear to show similar structure between Hawaii and the East Pacific Rise.) As noted above, model RG5.5 has a formal lateral resolution for structure of order ~1000 km (~10°) [Zhang and Tanimoto, 1993]. The fact that RG5.5 consistently shows hotspot anomalies is empirical evidence to us that this model is seeing lateral structure that is present at this scale in the asthenosphere. Vertical resolution is such that there appear to be

Figure 5. (a) Bathymetry as a function of age of Southern Atlantic Ocean seafloor for a tectonic corridor bounded by a northernmost flowline which abuts the Gough fracture zone, a southern flowline which abuts the Falklands escarpment, and isochrons at 100 Ma on the African and South American plates (Location is shown on Figure 4). Depths are from National Geophysical Data Center ship data and magnetic ages from Shaw and Cande [1990]. The dashed line shows the best fitting depth-√age curve for the African Plate (280 m/√Ma) shown as an "isostatic" reference depth for South American seafloor. South American seafloor is considerably deeper (up to ~1 km at 100 Ma) than this reference depth, suggesting that it is being dynamically supported by stresses from asthenosphere or mantle flow. (b) Geoid as a function of age for the same Southern Atlantic Ocean tectonic corridor as above. The dashed line shows the predicted geoid signal [Haxby and Turcotte, 1978] due to lithosphere cooling and subsiding at a rate of 280 m/√Ma. It fits well with the observed geoid on the African side of this corridor, suggesting that this geoid signal is predominantly due to (passive) thermal plate cooling with age. In contrast, the South American side has a ~20-m-large residual geoid signal which cannot be explained by the thermal structure of the lithosphere. (c) The ratio of residual geoid height Δn to residual depth Δd allows us, in the long wavelength limit, to infer a "geoid compensation" depth z of ~200 km for the denser mass anomaly which is leading to this anomalous geoid signal. In this cartoon the heavy solid line shows predicted geoid signal from cooling lithosphere. Heavy dashed line shows predicted depth from √age-cooling of the lithosphere. Observed geoid is shown by a thin solid line, while observed seafloor depth is sketched as the top of the lithosphere. Densities of water, asthenosphere, lithosphere, and mesosphere are shown as ρw, ρa, ρl, and ρm, respectively. An attractive hypothesis for the origin of this compensating mass anomaly is that the same asthenosphere pressure variation that leads to a dynamic depression of the seafloor also leads to uplift of the underlying asthenosphere-mesosphere density horizon. This can occur if the asthenosphere is less dense as well as more fluid than underlying mesosphere. The equal and opposite mass anomalies at the top and base of the asthenosphere to compensate the pressure variations due to asthenosphere flow are an effect that we call "dynamic isostasy". (d) Asthenosphere flow induced by the westward migration of the South American continent over the mantle mesosphere would produce a pressure drop that would lead to the depression of the seafloor [Phipps Morgan and Smith, 1992] and a concurrent elevation of an asthenosphere-mesosphere boundary if the asthenosphere is lighter than underlying mesosphere. This situation could arise if the asthenosphere is hotter and/or has had more basalt extracted from it during episodes of melting. In this cross section the arrow-tail beneath the ridge axis indicates that the asthenosphere in this cross section was fed by (out-of-plane) along-axis flow. Arrows show the absolute motions of the asthenosphere, lithosphere, and continents in this system.
Residual Depth/Geoid Evidence for a Large, Dynamic, Asthenospheric Geoid Signal

Residual seafloor depth anomalies are the surface dynamic relief caused by viscous stresses arising from mantle flow. Joint analysis of these depth anomalies and their corresponding geoid anomalies let one determine how deep the density anomalies are that compensate, through viscous flow-induced stresses, the deficit or excess surface mass anomaly associated with the dynamic deflection of the seafloor. Because the surface geoid anomaly reflects both the surface mass deflection and the compensating mass anomaly, the long-wavelength relationship between the dynamic depth anomaly $\Delta d$ and the dynamic geoid anomaly $\Delta n$ yields the depth $z$ of these compensating mass anomalies: $z = (\Delta n / \Delta d) g / (2\pi G (\rho_l - \rho_w))$ [Haxby and Turcotte, 1978]. A shallow $\sim 150$-300-km depth $z$ would imply dynamic asthenosphere support for these mass anomalies while a deeper depth would imply that these anomalies reflect stresses due to mantle flow driven by deeper-lying mass anomalies.

This depth/geoid experiment has been typically used in marine geophysics to determine the compensation depth of oceanic plateaus [Sandwell and MacKenzie, 1989] and hotspot swells [McNutt, 1988, Sandwell and MacKenzie, 1989]. The usual difficulty in this exercise lies in determining what parts of the seafloor depth and geoid anomalies are due to viscous flow instead of the “isostatic” effects of crustal thickness variations, sediment loading, or of different lithosphere thermal structures. If we examine two conjugate ridge flanks that share a common ridge origin, then portions of the seafloor of the same age are most likely to be underlain by similar thickness crust and litho-
sphere that reflects their common ridge thermal structure. If we examine conjugate ridge flanks for an accretion corridor which has experienced little off-axis or hotspot volcanism on either flank then these effects, too, will be relatively small. We also look for sites where both flanks are at the same latitude so that they have experienced similar sedimentation histories after they left the spreading center.

These stringent criteria restrict our initial analysis to a single corridor in the South Atlantic (Figure 5a) which also has good magnetic basement age determinations [Shaw and Cande, 1990] and a relatively simple and well understood spreading history [Shaw and Cande, 1990]. We truncate this corridor at 100 Ma on each side before it reaches major sediment basins near the African or South American continental margins. (Note that the sediment correction is not critical here. Mean sediment thicknesses are less than 500 m, and the differences in sediment thickness between seafloor of similar age on conjugate flanks are less than 200 m throughout our initial study corridor [Nürnberg, 1988].) The South American plate moves faster over the mantle than its conjugate African plate (~30 mm/yr versus 10 mm/yr [Morgan, 1981]). It also subsides more rapidly with age, as predicted by a model for asthenosphere flow induced topography [Phipps Morgan and Smith, 1992]. Since the African plate moves much more slowly over the mesosphere than the South American plate, we can use the African flank of the conjugate margin corridor to calibrate the seafloor depth and geoid effects of lithosphere cooling and growing with age. We find an African thermal subsidence rate of 280 m/My (Figure 5b). The predicted geoid slope due to thermal subsidence [Haspy and Turtcoit, 1978] is 0.14 m/My, in reasonably good agreement with the observed African geoid slope of 0.10 m/My (see Figure 5c).

The resulting dynamic depth and geoid anomalies of the South American plate at 100 Ma seafloor are 1 km and 20 m, respectively (Figure 5), which implies a compensating negative (buoyant) mass anomaly centered at ~200 km depth. This shallow depth of compensation is intriguing because it would not a priori be predicted by a model of dynamic surface deflections due to deep mantle mass anomalies (e.g. the subducting Nazca plate), but would be a consequence of asthenosphere flow if the asthenosphere is both less dense as well as less viscous than underlying mantle. The same asthenosphere pressure drop which will depress the South American seafloor behind the westward migrating South American continent can elevate the asthenosphere-mesosphere interface as a quasi-static effect which we call "dynamic isostasy" (Figure 5d). The resulting compensation depth of ~200 km could be the base of the asthenosphere where a buoyantly stable thermal stratification exists. A 200°C temperature change at this level between a plume-fed hot asthenosphere and cooler mesosphere would result in both a 1% reduction in density and a correlated viscosity reduction of several orders of magnitude as discussed earlier.

Global Geoid Evidence for Asthenosphere Flow

We see in Figure 5 that the geoid appears to be much more sensitive to asthenosphere flow than does the depth of the seafloor. If we examine conjugate ridge flanks for an accretion corridor which has experienced little off-axis or hotspot volcanism on either flank then these effects, too, will be relatively small. We also look for sites where both flanks are at the same latitude so that they have experienced similar sedimentation histories after they left the spreading center.

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Global Geoid Evidence for Asthenosphere Flow

We see in Figure 5 that the geoid appears to be much more sensitive to asthenosphere flow than does the depth of the seafloor. This would suggest that the global oceanic geoid (Plate 1) should at least qualitatively match the predictions of the pattern of asthenosphere flow that we envision. We think it does. Analysis of an idealized (one-dimensional) asthenosphere flow problem implies that partial asthenosphere return flow from subduction zones towards ridges will lead to geoid highs (as well as shallower seafloor depths) associated with convergent plate mar-
Plate 1. The oceanic geoid, cosine tapered between degrees 12-20 to remove short wavelength structure, is shown by three Lambert equal-area projections centered about the Atlantic, Pacific, and Indian Ocean Basins. Hotspots are shown by red triangles. Apart from 12 or so major hotspots, any hotspot list is fairly subjective. Our list is based, with slight modifications, on the list by Richards et al. [1988]. In particular, we have added the Rurutu [Smith, et al., 1989], Rarotonga [Smith, et al., 1989], Balleny [Green, 1992], and Louisville [Lonsdale, 1988] hotspots, and removed five “hotspots” that we feel are likely to be (minor) volcanism. The geoid is positive near the subducting Pacific, Indian, and Nazca plates, which we propose reflects the pressure elevation that drives limited asthenosphere return flow from subduction zones towards oceanic spreading centers. The geoid is also elevated over major hotspot concentrations in the southwest Indian, northern and eastern Atlantic, and south-central Pacific Oceans, which we propose reflects the pressure high that drives asthenosphere flow away from hotspot sources (as well as a correlated higher “young” asthenosphere temperature near the hotspot sources of asthenosphere). The geoid is negative behind migrating passive continental margins. We think this reflects the pressure drop needed for asthenosphere flow into the retreating continental wake. See text for further discussion.
Implications for Large-Scale Mantle Flow and Global Volcanism

The existence of a weak and plume-fed suboceanic asthenosphere may profoundly shape the structure of mantle convection. In particular, it means that plate motions and the positive buoyancy force of old, cold oceanic lithosphere only directly interact with deeper mantle flow at subduction zones where the lithosphere penetrates beneath the asthenosphere layer. This different paradigm leads to several interesting hypotheses for both large-scale and smaller-scale aspects of mantle flow and melting patterns.

Implications for Plate Motions

The forces which drive plate motions are most likely associated with the subduction of cold, dense slabs and the resistance and interactions of plate subduction with deeper mantle flow [Forsyth and Uyeda, 1975; Chapple and Tullis, 1977], and possibly the interaction of continental cratons with underlying high viscosity mantle (cf. Figure 1b). For present-day plates which move rapidly over the lower mantle (Indian, Pacific, Nazca, Cocos), the major driving and resisting forces would be associated with slab buoyancy and resistance to slab penetration since the continental craton area is small with respect to the area of the subducted slab section of the plate that lies between ~300-700 km depth (a minimum estimate for the depth interval where the subducting slab would interact with high viscosity mantle which opposes slab penetration).

However, asthenosphere drag still may be an important contributor (25-50%) to the forces resisting plate motion over the mesosphere. In this case the thickness of the asthenosphere layer could play an important role in modulating the “speed limit” for an oceanic plate. The resistance to differential plate versus mesosphere motion across a narrow asthenosphere channel will depend linearly on viscosity but to the inverse third power of the channel thickness. Hence small changes in channel thickness can dramatically change the shear stress generated by asthenosphere shear. The balance between the rate of plume supply of new asthenosphere and lithosphere growth that consumes asthenosphere could lead to a quasi-steady plate speed. For example, faster plate speeds would subduct more lithosphere which would consume more asthenosphere. In time a greater rate of asthenosphere consumption than the rate of supply of new asthenosphere by hotspot upwelling would lead to a thinner asthenosphere with a correspondingly greater drag to plate motion. The plate would slow until plume supply and plate removal were in balance.

A major increase in the absolute speed of the Indo-Australian plate occurred at the same time as a major episode of basaltic flood volcanism [Larson, 1991]. Both may be associated with the arrival of a plume “head” which leads to a rapid change in asthenosphere thickness (as well as flood basalts [Morgan, 1981]) and hence a rapid drop in asthenosphere resistance to plate motion. This mechanism also predicts that changes in a plate’s velocity should be correlated with the emergence (or submergence) of the passive continental margins which are part of that plate. The asthenosphere thickening which leads to faster plate motions could lead to a smaller deep seafloor depression behind the migrating continent. Faster absolute plate motions in the absence of asthenosphere thickening could lead to dynamic submergence of the retreating passive continental margin or the emergence of an advancing continental margin [Phipps Morgan and Smith, 1992].

Implications for Large-Scale Mantle Structure and Flow

The Pacific Ocean asthenosphere should be regarded as a stress-free as opposed to a moving lithospheric boundary condition on underlying mantle flow. Circum-Pacific subduction zones are then likely to drive diffuse lower-mantle upwelling more centrally sited within the Pacific basin (cf. Figure 1b), in accord with seismic observations of lower mantle structure [Su and Dziewonski, 1991; Tanimoto, 1990; G. Masters and H. Bolton, Long-period S travel times and the three-dimensional structure of the mantle, submitted to Journal of Geophysical Research, 1994].

If plumes feed all asthenosphere (whose cooling is the source for oceanic lithosphere), there may be no upwelling outside of plumes and, instead, a mantle “rain” towards a hot, weak D" CMB layer where material heats and flows laterally towards rising plumes of D" material (cf. Figure 1c). This pattern of flow could override the “natural” thermal convection due to internal mantle heating. Heat would still be lost to lithosphere growth at the surface and lithosphere resorption into lower mantle flow but it would not determine the resulting form of the mantle flow pattern which would be controlled by the smaller heat loss out of the core.

The D"-to-plume-to-asthenosphere-to-lithosphere rheological boundary layer structure could easily lead to a hierarchical form of whole mantle convection which involves two weakly interacting flow patterns. Narrow pipes of plume upwelling are akin to structures seen in very high Rayleigh-number (constant viscosity) convection. Slow, large scale background mantle flow due to lithosphere subduction is akin to the structures seen in low Rayleigh-number (constant viscosity) convection experiments and the “red” thermal structure observed in three-dimensional tomographic inversions of the seismic velocity structure of the lower mantle [Su and Dziewonski, 1991]. The mantle may appear “red” to seismic tomographic studies because they only image the slow, large-scale component of mantle flow, while not imaging a large plume component of mantle upwelling.

Implications for Global Heat Flow

A long standing apparent paradox in global heat flow is the fact that the mean continental and oceanic heat flow are quite similar in spite of a much greater concentration of heat-producing radioisotopes in continental crust. If corrected for radioactive crustal heat production, mantle heat flows into the base of continental crust at roughly half the rate it flows into the base of oceanic crust [Sciater et al., 1981]. This is readily explained if the source temperature beneath continents is lower (mesosphere temperature ~10-20% lower than asthenosphere temperature) and the conductive lid is thicker (perhaps twice as thick because of a compositionally lighter continental lithosphere density [Jordan, 1981]) than for old oceanic lithosphere.

Implications for Off-Ridge Volcanism

Patterns of off-axis volcanism may be related to ridgeward asthenosphere flow from off-axis plume sources. In general, mantle melting is thought to be a “passive” pressure release process. Thus relatively deep (sublithosphere) melting could occur at off-axis plumes while predominantly shallower melting would occur as asthenosphere melts under a mid-ocean ridge where the axial lithosphere thickness is extremely shallow. As asthenosphere moves from plume source to near-ridge sink it will move under progressively thinner and younger lithosphere. This would impart a (small) vertical component to asthenosphere motions which
should result in small amounts of asthenosphere melting and off-axis volcanism. In contrast, where asthenosphere flows in the direction of plate motions, the asthenosphere-lithosphere boundary will progressively deepen, which should lead to no pressure release melting. This mechanism would predict a strong correlation between the abundance of off-axis seamount production and the geometry of plume sources and ridge sinks. In particular it would predict more volcanism on the side of a ridge which is closest to a nearby hotspot; a prediction which can be tested with current high-resolution Geosat altimetry coverage south of 30°S and forthcoming global ERS 1 high-resolution altimetry coverage.

Implications for Upper Mantle Seismic Anisotropy

Seismic anisotropy is likely to be strongest within the asthenosphere where large horizontal shear accompanies asthenosphere flow and within the oceanic lithosphere which forms from the cooling and accretion of underlying asthenosphere. Anisotropy may be most pronounced near diverging plume sources; regions near subduction zones where asthenosphere return flow occurs [Phipps Morgan and Smith, 1992], and beneath mid-ocean ridges; all are likely to be the regions of greatest horizontal or vertical asthenosphere shear.

Summary

We think that it is time to critically re-examine the dynamic implications of a weak asthenosphere layer on mantle flow, considering, in particular, the effects of a strong difference in lateral strength between continental roots and an asthenosphere layer beneath the oceanic lithosphere. We suggest that this asthenosphere layer, fed by plumes and consumed by lithosphere accretion and plate subduction, may offer a simple explanation for the existence of “passive” ridges and the “small” contribution to global heat flow of hotspot volcanism. The pattern of suboceanic asthenosphere flow may be able to be mapped through its dynamic and thermal effects on second order variations of ocean depth with age that are not explained by lithosphere cooling and growth [Phipps Morgan and Smith, 1992].

This hypothesis offers straightforward explanations for many of the current “paradoxes” that are emerging as we learn more about the oceanic seafloor; for example, the origin for asymmetric seafloor subsidence about a spreading center, or for long wavelength variations in the height of the ridge that are not explained by simple plate cooling models. It is also qualitatively consistent with the shape of the geoid, and with emerging seismic measurements of the structure of the oceanic upper mantle. The primary tests for this paradigm will come from several areas of geophysical study. Better seismic pictures of the seismic velocity and anisotropy structure of the oceanic upper mantle, more knowledge of the patterns of seafloor depth anomalies and seafloor volcanism (and geochemical relationships between volcanism at hotspots, ridges, and intervening seafloor), and a better understanding of the effects of shallow lateral viscosity variation on global mantle flow should, in concert, allow workers to more fully test this paradigm in the upcoming decade. We hope that this paper serves as a helpful stimulus to re-examine the roots of currently preferred paradigms for mantle flow.

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