On the shallow origin of hotspots and the westward drift of the lithosphere

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ABSTRACT

Intraplate migrating hotspots, which are unrelated to rifts or plate margins in general, regardless of their origin in the mantle column, indicate relative motion between the lithosphere and the underlying mantle in which the hotspot source is located. Pacific plate hotspots are sufficiently fixed relative to one another to represent an independent reference frame to compute plate motions. However, the interpretation of the middle asthenosphere rather than the deep lower mantle as the source for intraplate Pacific hotspots has several implications. First, decoupling between the lithosphere and subasthenospheric mantle is greater than recorded by hotspot volcanic tracks (>100 mm/yr) due to undetectable shear in the lower asthenosphere below the magmatic source. The shallower the source, the larger the décollement.

Second, computation of the westward drift is linked to the Pacific plate and assumes that the deep lower mantle, below the decoupling zone, sources the hotspots above. The Pacific plate is the fastest plate in the hotspot reference frame and dominates the net rotation of the lithosphere. Therefore, if decoupling with the subasthenospheric mantle is larger, the global westward drift of the lithosphere must be faster than present estimates, and may possibly vary between 50 and 90 mm/yr. In this case, all plates, albeit moving at different velocities, move westward relative to the subasthenospheric mantle.

Finally, faster decoupling can generate more shear heating in the asthenosphere (even >100 °C). This amount of heating, in an undepleted mantle, could trigger scattered intraplate Pacific volcanism itself if the viscosity of the asthenosphere is locally higher than normal. The Emperor-Hawaiian bend can be reproduced when bent viscosity anisotropy in the asthenosphere is included. Variations in depth and geometry in the asthenosphere of these regions of higher viscosity could account for the irregular migration and velocities of surface volcanic tracks. This type of volcanic chain has different kinematic and magmatic origins from the Atlantic hotspots or wetspots.

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which migrate with or close to the oceanic spreading center and are therefore plate-margin related.

**Keywords:** shallow hotspots, shear heating, asthenosphere viscosity, westward drift, reference frame

**INTRODUCTION**

Because Hawaii is the archetypal hotspot, the depth at which it is sourced is particularly important, not least because its volcanic track constrains the computation of the absolute plate motions on Earth and it influences greatly models of mantle dynamics. If the origin of Pacific intraplate hotspots is in the asthenosphere, what triggers the extraction of the extra melt? Why are the hotspots isolated, scattered, and sporadic? Their tracks suggest relatively stationary sources in the mantle (Wilson, 1963, 1973; Morgan, 1971) beneath an overlying lithosphere that moves independently. In other words the mantle moves east-southeast along the Hawaiian track (e.g., Ihinger, 1995). This motion requires a decoupling between the two layers, and the zone of decoupling is usually thought to be located in the asthenosphere; that is, the layer with the lowest viscosity in the mantle (e.g., Anderson, 1989). This decoupling zone is too often disregarded in geodynamic studies.

The nature and depth of hotspots is a matter of debate (Anderson, 2000; Courtillot et al., 2003). Hotspots form in most geodynamic environments—along or close to oceanic ridges and continental rifts (e.g., Iceland, Afar), along transform faults (e.g., Réunion), in the hanging wall of subduction zones (e.g., Yellowstone), and in intraplate settings (e.g., Hawaii). Therefore, different hotspots may have different depths of origin (Doglioni, 1993). Geochemical, geophysical, and petrological evidence is growing for a shallow source for hotspots (Bonatti, 1990; Anderson, 1999; Smith and Lewis, 1999; Foulger, 2002). The depth of origin of Hawaiian magmatism and similar Pacific intraplate hotspots is also strongly debated. A few authors (Smith, 1993; Smith and Lewis, 1999; Anderson, 2000; Janney et al., 2000; Green et al., 2001), argued against the deep mantle plume paradigm (e.g., Schubert et al., 2001). Some authors also propose crack propagation in the lithosphere along hotspot tracks as the reason for Pacific intraplate magmatism (Winterer and Sandwell, 1987; Sandwell et al., 1995; Lynch, 1999). Moreover, some seismic tomography results do not show lowered velocity beneath the Hawaiian hotspot (Wolfe et al., 2002; Foulger et al., 2003). Many workers still prefer a deep source at the core-mantle boundary (CMB) for some hotspots (Zhao, 2001; Courtillot et al., 2003), although there are now fewer adherents to this proposal than in the past (Ritsma and Allen, 2003). Heatflow does not show any significant anomaly on the Hawaiian rise (Shaw and Jackson, 1973; von Herzen et al., 1989; Stein and Stein, 1992, 1993). Deep mantle plumes would be expected to generate a long-wavelength regional increase in heatflow values (Anderson, 2000). The largest heat dissipation of the mantle occurs along mid-ocean ridges, and hotspots represent the smallest component of the Earth’s cooling budget.

Hawaiian lavas are thought to come from ~80–120 km depth or shallower. Petrological estimates of the temperature anomalies beneath Hawaii compared with ridges vary from zero to a maximum of ~200 °C (Green et al., 2001; Gudfinnsson and Presnall, 2002). Plume models require temperature anomalies of 200–600 °C (e.g., Cordery et al., 1997). A small change in temperature, volatile content, or fertility of the upper mantle can lead to a large change in the extent of partial melting and melt volumes (Yaxley, 2000; Green et al., 2001; Asimow and Langmuir, 2003; Foulger et al., 2003). A number of authors (Jackson et al., 1972; Shaw and Jackson, 1973; Shaw et al., 1980) noted a periodicity of the magmatism in the Hawaiian-Emperor chain and its en echelon distribution of volcanoes, and they emphasized the role of the stress field in controlling magma generation. Shaw (1973) and Shaw and Jackson (1973) proposed that linear island chains in the Pacific are related to mechanical rather than thermal anomalies. Three types and depths of mantle source were proposed by Courtillot et al. (2003) in an attempt to reconcile the different opinions about the origins of hotspots.

In this paper, we explore the kinematic and magmatic implications and consequences of shallow interpretations of hotspot sources. We show that the shallower the source of the magmatism, the faster the westnorthwest motion of the Pacific and the entire Earth’s lithosphere relative to the subasthenospheric mantle. Moreover, the decoupling between lithosphere and subasthenospheric mantle should produce shear heating in the intervening asthenosphere sufficient to generate locally intraplate Pacific magmatism (e.g., at Hawaii).

**ASTHENOSPHERIC SOURCE OF PACIFIC HOTSPOTS AND WESTWARD DRIFT OF THE LITHOSPHERE**

The Nuvel 1 (DeMets et al., 1990) and the Space Geodesy ITRF2000 and NASA database models (Hehlin et al., 2003) describe plate motions. They are based on the artificially imposed assumption of a no-net-rotation (NNR) reference frame: Plates are imagined to move relative to a fixed Earth center and the lithosphere is stationary relative to the underlying mantle, or the sum of its movements is zero. However, we know that the lithosphere moves relative to the subasthenospheric mantle, as not only suggested by the Hawaiian and similar tracks, but as kinematically required by the relative migration of plate margins.
The ITRF2000 model (e.g., Heflin et al., 2003) is excellent for describing relative plate motions and is also considered an absolute reference frame (relative to the Global Positioning System constellation or the Earth’s center of mass). However, when magmatic sources are included in the kinematic analysis, the movement of the lithosphere relative to the mantle should then be taken into account in an “absolute” plate motion analysis, and the NNR should be abandoned. The absolute reference frame is quite a controversial issue. Typically, the absolute reference frame is considered to be that of the hotspots.

When plate motions are measured in the hotspot reference frame, the lithosphere shows a net westward drift (Bostrom, 1971; O’Connell et al., 1991; Ricard et al., 1991). This westward drift persists when plate motions are computed relative to Antarctica (Le Pichon, 1968; Knopoff and Leeds, 1972). Usually the drift is computed using a larger number of hotspots on other plates, where the net rotation of the lithosphere is 49 mm/yr (0.44 ± 0.11 degrees/m.y.) about a pole at 56° S, 70° E (Gripp and Gordon, 2002). Most of the hotspots used, however, are not fixed; nor do they represent a fixed reference frame, because they are located on plate margins, such as moving ridges (Galapagos, Easter Island, Iceland, Ascension), transform faults (Réunion), above subduction zones (Yellowstone), or in continental rifts (Afar), all features that move relative to one another and to the mantle. To know whether the hotspots are fixed relative to one another (Molnar and Stock, 1987; Steinberger, 2002) is crucial for determining a reliable hotspot reference frame. Norton (2000) grouped hotspots into three main families that have very little internal relative motion (Pacific, Indo-Atlantic, and Iceland). He concluded that a global hotspot reference frame is not possible because Pacific hotspots move relative to Indo-Atlantic hotspots and to Iceland. However, in his analysis, Pacific hotspots are reasonably fixed relative to one another during the past 80 Ma. Present-day plate motions, based on space geodesy data (Heflin et al., 2003), confirm that Pacific intraplate hotspots move as a unit, whereas all other so-called “hotspots” move in various directions and with different velocities.

The westnorthwest motion of the Pacific plate relative to the underlying mantle is inferred from the Hawaiian and other major intraplate hotspot tracks (Marquesas, Society, Pitcairn, Samoan, MacDonald), which show an average velocity of ~103–118 mm/yr. They also move along the same trend (290°–300° westnorthwest) and therefore are the only hotspots that appear to be fixed relative to one another, representing an apparently reliable reference frame for absolute plate motions. When the westward drift is computed only taking those Pacific intraplate hotspot tracks that are relatively fixed during the past few million years (Gripp and Gordon, 2002), the net rotation of the lithosphere rises to ~90 mm/yr (M. Cuffaro, 2004, personal commun.), with a pole of rotation still deviated with respect to the Earth’s axis of rotation.

Provided that the velocity of the Pacific lithosphere, \( V_p \), toward the eastsoutheast (110°–120°) is slower than that of the underlying subasthenospheric mantle, \( V_m \) (i.e., \( V_m > V_p \)), the relative motion corresponding to the westnorthwest delay of the lithosphere is \( V_m - V_p = 103 \) mm/yr. However, assuming that shear is distributed throughout the asthenospheric channel, a location for the Hawaiian melting spot in the asthenosphere instead of in the upper mesosphere would imply a greater displacement between the lithosphere and mesosphere than is recorded in hotspot traces (Fig. 1). Therefore, if Hawaiian magmatism is in the upper part of the asthenosphere, there should be a deeper component of shear, which would increase the total relative velocity. Greater velocity has two basic consequences: (1) It increases the global westward drift of the lithosphere, and (2) it increases shear heating in the asthenosphere. If the source of Pacific hotspots is in the middle of the asthenosphere, half of the lithosphere–subasthenospheric mantle relative motion is unrecorded, which means that the total relative displacement would amount to ~200 mm/yr. Using this velocity to compute the net rotation of the lithosphere toward the west, no one plate still moves eastward relative to the mantle.

There is evidence that the propagation rate of Pacific “hotspots” or seamount tracks has varied over time and even included jumps back and forth and oblique propagation relative to absolute plate motion. This evidence casts doubt on both the notion of absolute plate motions computed in the hotspot reference frame, and the nature of the magmatism itself. Different models have been proposed on the origin of these magmatic outpourings, such as deep plume, or alternatively, shallow plumes generated by cracks, or boudins of the lithosphere (Winterer and Sandwell, 1987; Sandwell et al., 1995; Lynch, 1999; Natland and Winterer, 2003). Compositional heterogeneity in the mantle has also been invoked, particularly because no significant positive thermal anomaly relative to normal mid-ocean ridge has been observed (e.g., Stein and Stein, 1992, 1993). Janney et al. attribute a velocity for the Pukapuka volcanic ridge (interpreted as either a hotspot track or a leaky fracture zone) in the eastern central Pacific of ~200–300 mm/yr (see Fig. 2a in Janney et al., 2000), with a ridge age of between 5 and 12 Ma. They also inferred a shallow mantle source for Pacific hotspots, based on their geochemical characteristics.

**SHEAR HEATING AS A CAUSE FOR PACIFIC HOTSPOTS?**

Kennedy et al. (2002) have shown how mantle xenoliths recorded shearing possibly located at the lithosphere-asthenosphere interface. This shearing supports the notion of flow in the upper mantle and some decoupling at the base of the lithosphere (Russo and Silver, 1996; Doglioni et al., 1999; Bokelmann and Silver, 2000). The fastest plate on Earth in the hotspot reference frame (i.e., the Pacific) is the one affected by the most widespread intraplate magmatism. Is there a relation between these two observations? In Africa, so-called “hotspots” are located along rift zones, and therefore they are not simple intraplate or boundary-free magmatic sources. Because the Pacific appears to be the fastest-moving plate relative to the mantle (e.g., Gripp...
and Gordon, 2002), in the asthenospheric decoupling zone at its base there should occur the highest shear or viscous heating. This heating is a function of the lithostatic load acting on the decoupling layer, the friction and velocity of the decoupling (shear stress), the viscosity of the medium, the velocity of the decoupling, the thickness and thermal gradient between the top and bottom of the decoupling zone (i.e., the asthenosphere), thermal diffusivity, specific heat, and so on.

The ability of frictional heating to cause partial melting, particularly in subduction zones, remains a matter of controversy (Schubert et al., 2001). Yuen et al. (1978) studied frictional heating on slip zones using thermally activated creep laws for the deformation. Because rock becomes weak as the solidus is approached, these authors concluded that the maximum temperatures associated with frictional heating are always less than those required for partial melting. Shaw (1969, 1973) suggested shear heating in the asthenosphere generated by the decoupling between lithosphere and subasthenospheric mantle related to tidal torque (Shaw et al., 1971) as the cause of Hawaiian magmatism. With a different mechanism, such as a combination of lithospheric flexure and viscous shear heating at the base of the lithosphere, Hawkins and Natland (1975) and Natland (1980) proposed a model of magma generation unrelated to a deep mantle plume for the Samoan linear volcanic chain.

It is noteworthy that the fast Pacific plate overlies asthenosphere with the lowest viscosity ($5 \times 10^{17}$ Pa·s; Pollitz et al., 1998), and possibly the least depleted mantle, which is therefore the most prone to melt. Because of the melting characteristics of peridotite with minor carbon + hydrogen (the lherzolite–[C + H+O] system), the asthenosphere is already partly molten (e.g., Schubert et al., 2001) and is at a temperature of $\approx 1430$ °C (e.g., Green and Falloon, 1998; Green et al., 2001). A rise in temperature of only a few tens of degrees will increase the degree of melting, which, in a deforming material, will migrate toward the surface. We postulate that locally, the viscosity of the asthenosphere can also increase (e.g., $10^{19}$ Pa·s due to refractory geochemical anisotropy), or decrease due to locally higher water activity. Shear stress could be irregularly distributed in such inhomogeneous material and consequently, higher shear heating may develop locally.

Assuming a lithosphere 10,000 km long, 100 km thick, and with a velocity of 200 mm/yr relative to the subasthenospheric mantle, can the internal shear or viscous heating generate sufficient excess temperature to produce a large amount of melt in the low-velocity channel? Watson and McKenzie (1991) estimated that for Hawaiian magmatism, the maximum mantle adiabat has a potential mantle temperature of 1830 K, the degree of partial melting is $\approx 6.9\%$, and the depth range of melting is 72–127 km (i.e., in the asthenosphere). Recently, Sisson (2003) suggested that trace and major element features of Hawaiian tholeiites point to separation from sources shallow in the garnet lherzolite stability field, at a pressure of $30 \pm 3$ kb, corresponding to depths
of ~90–100 km. In the same study, a temperature of 1470–1520 °C (~1740–1790 K) was suggested for the Hawaiian source.

Subsidence related to cooling of the oceanic lithosphere does not fit with old oceanic crust (>80–100 Ma) where the bathymetry flattens (Nagihara et al., 1996). Flattened oceanic bathymetric profiles have been related to radiogenic heating in convecting mantle (Jarvis and Peltier, 1980) or asthenospheric flow (Phipps Morgan and Smith, 1992). The apparent constant lithosphere thickness at 100–80 Ma has also been attributed to the form of the peridotite–(C + H+O) solidus when it is influenced by pargasite (amphibole) stability at <2.8–3.0 GPa (Green and Liebermann, 1976). As an alternative hypothesis, extra heating of the lithosphere could be attributed to the shear at the base of the plate. The hypothesis of shear heating as a mechanism for generating more melting is tested in the following section. Areas of higher viscosity, acting as asperities in the asthenospheric decoupling zone, should generate local tension in the overlying lithosphere, resulting in thinning of the lithosphere and upwelling of underlying the asthenosphere.

**THERMAL MODEL**

Asthenospheric shear flow may be represented by simple Couette flow between plane parallel walls, as shown in Figure 2, assuming a simple linear rheology. The upper wall \((y = 0)\), which is the base of the lithosphere, has a velocity \(u = u_0\) and temperature \(T = T_0\). The lower wall \((y = h)\) is the top of the subasthenospheric mantle, which has a velocity \(u = 0\) and temperature \(T = T_1\). The velocity profile:

\[
u = u_0 \left( 1 - \frac{y}{h} \right).
\]

Figure 2. Relative motion of the lithosphere and the subasthenospheric mantle is imposed. Asthenospheric shear flow may be represented by simple Couette flow between plane-parallel walls (Turcotte and Schubert, 2002), and the shear stress is assumed to be equal throughout the asthenospheric section. The upper wall \((y = 0)\), which is the base of the lithosphere, has velocity \(u_0\) and temperature \(T_0\). The lower wall \((y = h)\) is the top of the subasthenospheric mantle and has velocity \(u = 0\) and temperature \(T_1\).

is produced by the shear stress:

\[
\tau = \mu \frac{du}{dy} = \frac{\mu u_0}{h} = \tau_0,
\]

where \(m\) is the viscosity, \(u_0\) is the velocity of the lithosphere relative to the mantle, \(h\) the thickness of the asthenosphere, and \(y\) is the vertical axis. According to Turcotte and Schubert (2002), the rate of shear heating per unit volume is:

\[
\frac{\mu u_0^2}{h^2} = \text{constant},
\]

because the shear stress is constant and the velocity profile is linear.

The temperature distribution in the channel is given by the integration of:

\[
k \frac{d^2T}{dy^2} = -\frac{\mu u_0^2}{h^2},
\]

where \(k\) is the thermal conductivity, with the boundary conditions \(T = T_0\) at \(y = 0\), and \(T = T_1\) at \(y = h\). The solution in dimensionless form is (Turcotte and Schubert, 2002):

\[
\theta = \frac{T - T_0}{T_1 - T_0} = \frac{\mu u_0^2/2k}{h} \left( 1 + \frac{\mu u_0^2/2k}{h^2} \left( \frac{\mu u_0^2/2k}{h^2} \right) \right) - \frac{y^2}{h^2} \left( \frac{\mu u_0^2/2k}{h^2} \right),
\]

or

\[
\theta = \frac{y}{h} \left( 1 + \frac{PrE}{2} \right) - \frac{y^2}{h^2} \left( \frac{PrE}{2} \right),
\]

where

\[
E = -\frac{u_0^2}{C_p(T_1 - T_0)}
\]

is the Eckert number and \(C_p\) is the specific heat at constant pressure, and

\[
Pr = \frac{\mu}{\rho \alpha}
\]

is the Prandtl number, where \(\rho\) is the density and \(\alpha\) the thermal diffusivity.

The maximum excess temperature due to frictional heating, with respect to the linear temperature profile is (Turcotte and Schubert, 2002):

\[
\theta_{e}^{\text{max}} = \frac{PrE}{8},
\]

and occurs at \(y/h = 0.5\).
PETROLOGICAL MODEL

In major and minor element concentrations, mid-ocean ridge basalts (MORBs) are inferred to be derived from a well-mixed “pristine modern mantle” source, although detailed observations including isotopic signatures show that long-lived heterogeneity is required. Different isotopic end-members have been proposed. In the Hawaiian volcanic chain, differences both within and between volcanoes occur, including major element, minor element, trace element, and isotopic signatures. These differences have led to inferences of greater heterogeneity in this “hotspot” source. Particular trace element patterns and ratios and distinctive isotopic ratios have been attributed to primordial mantle, old oceanic crust, sediment, and arc signatures, implying complexity and longevity in the Hawaiian “hotspot” source and the absence of the homogenizing process that has given rise to the MORB source. The major element compositions of primitive magmas, and thus the nature of liquidus phases at various pressures and temperatures, have been interpreted as requiring a more refractory source and harzburgite residues for Hawaiian magmas compared to primitive MORB (Green and Falloon, 1998; Green et al., 2001). These petrological arguments have been linked to geophysical interpretations of the cause of the broad topographic high (the Hawaiian swell) underlying the Hawaiian volcanic chain to suggest that the buoyancy anomaly reflects a compositional rather than a thermal anomaly beneath Hawaii. The phenomenon of selective shear heating linked to viscosity variations in the asthenospheric mantle obviously influences these models.

The melting behavior of lherzolitic mantle (pyrolite) containing a little water and carbon (required for sources of either MORB or hotspot basalts) is defined by the peridotite-bearing lherzolite solidus to ~3 GPa and then by the garnet-lherzolite-(C + H$_2$O) solidus at deeper levels. The latter solidus is sensitive to mantle oxidation state (fO$_2$), which determines the CH$_4$/H$_2$O ratio of potential subsolidus fluid and thus controls solidus temperature (Green and Falloon, 1998). The asthenosphere is inferred to have lower viscosity because of a very small melt fraction—a region of incipient melting (Green and Liebermann, 1976; Green and Falloon, 1998). In Figure 4, the incipient melt region is contoured for melt fraction assuming 0.1% H$_2$O (and C content of ~0.01%, equivalent to 400 ppm CO$_2$), and the oceanic intraplate geotherm depicted lies between 0.5% and 1% melt contours. If shear stress and heating is distributed throughout the asthenosphere, a temperature increase of ~120 °C caused...
by shear heating would approximately double the melt fraction to 1.5% melting. If more localized compositional heterogeneity includes refractory, residual compositions lacking (H$_2$O + CO$_2$), then increased solidus temperature may produce higher viscosities, locally increased shear stress, and shear heating in the refractory volume. Alternatively, the fO$_2$ contrasts inferred at the interfaces between old subducted lithosphere suspended in the asthenosphere or in the subasthenospheric upper mantle, and normal MORB mantle (Green and Falloon, 1998) may focus both melting and strain in the boundary zone, augmenting melting by the shear-heating effect.

**KINEMATIC MODEL**

The Emperor-Hawaiian bend does not correspond to a change in plate motion of the Pacific plate because there is no corresponding bend in transform faults at the same time (Norton, 1995; Richards and Lithgow-Bertelloni, 1996; Raymond et al., 2000). Scholl (2003) noted that between ca. 76 and 43 Ma, the linear chain of Emperor seamounts records a southward-moving hotspot across a generally westnorthwest-moving Pacific plate. This observation means that the so-called “Emperor-Hawaiian bend” does not record a ~40° change (northnorthwest to northwest) motion of the Pacific plate but rather, a great slowing or cessation of hotspot drift. According to the same authors, while the Hawaiian chain was formed, the hotspot remained fixed with respect to the Earth’s spin axis. The Emperor-Hawaiian bend has been redated to ca. 47 Ma (Pilger and Handschumacher, 1981), and the redating of rocks still supports the picture that the volcanoes increase in age systematically to the northwest, away from the active volcanoes in Hawaii (Clague, 2003). Using the revised calibration of magnetic anomalies of Cande and Kent (1995), Norton (2000) pointed out an unexplained variation in plate motion at 47 Ma when comparing North and South America relative to the Pacific. Shaw et al. (1980) showed that the Emperor and Hawaiian volcanic chains must have the same magmatic origin, despite the bend. In other words, we can argue that the Emperor hotspot migrated obliquely to the Pacific absolute plate motion, whereas the Hawaiian chain is parallel to the Pacific absolute plate motion. How is it possible to reconcile these data in a realistic kinematic model?

Vertical and lateral movements of the magmatic source in the décollement zone (i.e., the asthenosphere) could account for these observations (Fig. 5). Anisotropy in the composition and viscosity of the asthenosphere, lying both oblique and parallel to the plate motion, can explain the Emperor and Hawaiian chains (Fig. 5). For an area of higher asthenospheric viscosity that is oriented northeast-southwest and generating melt, the overlying plate moving westnorthwestward will record a volcanic track oriented northwest-southeast, such as the Emperor chain (Fig. 5). In this way, the faster southward migration of the Emperor hotspot compared with Hawaiian migration rates can be explained because the oblique vector in a 10-m.y. incremental interval is longer (faster) than the plate motion vector (top panel of Fig. 5).

Tarduno et al. (2003) suggested a southward motion of the Emperor hotspot, based on paleomagnetic and radiometric data. Paleomagnetic data contribute to reconstructing palaeomagnetic latitude, but not longitude; therefore, the Emperor chain may also have had an eastward component, although smaller, as the Hawaiian track shows. In fact, as far as the westward drift is concerned, plate motions move with the mantle toward the east only a little slower. The southward component of Emperor hotspot
Figure 5. Kinematic map view model for the bending of the Emperor-Hawaiian chains, maintaining stable Pacific plate motion. The bend can be reproduced by introducing a bent, mirrorlike anisotropy in the décollement zone (i.e., the asthenosphere), where locally higher viscosity generates shear heating and related magmatism. The shaded area is a part of the Pacific plate progressively moving to the northwest and overriding the asthenosphere. The asthenosphere is assumed to have a typical viscosity of $10^{17}$ Pa·s apart from the gray line, where it rises to $\sim10^{20}$ Pa·s. In each panel, the red spot indicates the active magmatism. In the top panel, while the oblique Emperor seamount chain formed, the velocity of the southeastward-propagating volcanic ridge is faster because it is oblique to Pacific plate motion (the longer oblique arrow in the incremental 10-Ma interval).
migration is thus not necessarily related to motion of a deep lower mantle plume, because the surface kinematics can also be explained by a heterogeneous asthenosphere, in which a viscosity anomaly lies in the low-velocity layer, oblique to the absolute motion of the overlying plate (Fig. 6). Our model is clearly ad hoc in the sense that we infer anisotropy in the asthenosphere from the presently observable volcanic trail. Therefore the model needs to be tested. We postulate two unconstrained assumptions: (1) the Emperor chain formed obliquely to the 80- to 47-Ma Pacific absolute plate motion; and (2) there are similar anomalies in the Pacific asthenosphere that generated the Louisville, MacDonald, and Emperor-Hawaiian volcanic tracks. In this interpretation, the number of volcanic tracks would equal the number of anomalies in which the asthenosphere viscosity was and still is high enough to generate extra melt in the shear zone.

**LITHOSPHERE–SUBASTHENOSPHERIC MANTLE MOTIONS**

Relative plate motions can be estimated with great accuracy with space geodetic data (e.g., Robbins et al., 1993; Heffin et al., 2003), refining the earlier Nuvel 1 plate-motion model (DeMets et al., 1990). The east Pacific Rise (EPR), separating the Pacific and Nazca plates, opens at a rate of 128 mm/yr just south of the equator (e.g., Heflin et al., 2003). At the same latitude, the shortening along the Andean subduction zone, where the Nazca plate subducts beneath South America, has been computed to be ~68 mm/yr. These relative motions, when considered in a reference frame in which the Hawaiian hotspot is fixed and positioned in the subasthenospheric mantle, suggest that the Nazca plate moves eastward relative to the subasthenospheric mantle at ~25 mm/yr (Fig. 7, option 1). If we assume that the source of

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**Figure 6. The irregular distribution of an area of higher viscosity (e.g., \(10^{20-21}\) Pa·s) in the Pacific asthenosphere (viscosity, \(10^{17}\) Pa·s), such as that shown by the curved gray band in the pretransit cartoon, could account for the bend in the Emperor-Hawaiian track, in spite of a stable absolute Pacific plate motion direction. The pretransit geometry mirrors the resulting volcanic chain shown in the posttransit cartoon. X, Y, Z, and H in the asthenospheric asperity correspond to similarly labeled volcanoes in the surface trail. Red letters show the active magmatism. The gray volcanic track in the left panel indicates inferred, not yet existing volcanoes.
the Pacific intraplate hotspots is, instead, in the middle asthenosphere and that half of the relative motion of the lithosphere–subasthenospheric mantle is missing in the Hawaiian track (Fig. 1), the rate of motion could be as high as 200 mm/yr, as is also suggested by some segments of the Pukapuka volcanic ridge (Janney et al., 2000). Greater speeds, as the Pukapuka ridge suggests, implies a source from a deeper section of the asthenosphere relative to the Hawaiian source (Fig. 1). In this model, the Nazca plate would instead move west relative to the mantle at 72 mm/yr (Fig. 7, option 2), and therefore all three plates would move westward relative to the subasthenospheric mantle. This scenario is in agreement with shear-wave splitting observations (Russo and Silver, 1994) and the gentle dip of the Andean slab, which both suggest relative eastward mantle flow. Similar eastward mantle flow was proposed for the North American plate (Silver and Holt, 2002). The low dip of the Andean slab has alternatively been attributed to the age of the subducting lithosphere. However, the age has been shown to be insufficient to explain the asymmetry between westerly directed (steep and deep) versus easterly directed (low dip and shallow) subduction zones. In fact the geographically related asymmetry persists even where the same lithosphere (regardless of whether it is oceanic or continental) subducts on both sides, such as in the Mediterranean orogens (Doglioni et al., 1999). In western Central America, the dip of the easterly directed slab is steeper in spite of the young Cenozoic age of the downgoing oceanic Cocos lithosphere. However, it is known to be shallow, because most of the seismicity stops at ~200 km, and its dip can be explained by oblique-lateral subduction and by the continental lithospheric thickness of the upper plate. Another counterexample is the westerly directed Sandwich subduction zone in the southern Atlantic, where the Cenozoic-to-zero age oceanic lithosphere dips very steeply and penetrates deep as in the Barbados or Marianas subduction zones.

Jordan (1974) questioned westward drift of the lithosphere on the grounds that the viscosity required to allow decoupling of the lithosphere is too low (10^{11} Pa s) compared with the known viscosity of the asthenosphere. However, new mechanical models have subsequently been developed to explain the net rotation of the lithosphere with even higher viscosity values when tidal drag is coupled with the opposite torque generated by downwelling of heavier materials in the core and mantle and with asymmetric mantle convection (Scoppola et al., 2005). Another relevant point raised by Jordan (1974) is that the westward drift at that time computed by Minster et al. (1974) of 12 mm/yr was too low compared to the eastward component of a number of plates (e.g., Cocos and Nazca moving eastward at rates of >70 mm/yr). This rate would imply that the spreading rate of the EPR is faster than the absolute westnorthwest velocity of the Pacific plate, based on the Hawaiian volcanic track, and the Nazca plate would move relatively eastward faster than the subasthenospheric mantle (Fig. 7, option 2). This faster speed than the spreading rate of the EPR, with the Nazca plate also moving westward relative to the subasthenospheric mantle (Fig. 7, option 2).

The lower the asthenosphere viscosity, the faster plates move westward; that is, there is low coupling between lithosphere and subasthenospheric mantle. So why does the Nazca plate move slower with respect to the Pacific plate if it shifts west and travels above asthenosphere, which was previously underneath the Pacific, allowing fast motion? In other words, when modeling plate tectonics in the frame of westward drift of the lithosphere, one unsolved issue is why the eastern plate moves slower relative to the western plate, if velocity is controlled by lithosphere-asthenosphere coupling, which should continuously vary with the westward motion of the lithosphere. Braun et al. (2000) show that water extraction during melting results in higher viscosity in the residual mantle of up to two orders of magnitude. The mantle, once depleted beneath the ridge, should be cooler, less dense, and more viscous. Therefore, eastward relative movement of the mantle underneath the ridge could explain the higher seismic velocities detected by mantle tomography in the Nazca asthenosphere relative to the Pacific asthenosphere in the 140–300 km interval by Romanowicz and Gung (2002), as well as the higher Rayleigh and Love wave phase velocities in the asthenosphere beneath the eastern side of most of the Atlantic Ocean compared to its western side (Silveira et al., 1998; Silveira and Stutzmann, 2002).

Our model assumes a fixed subasthenospheric mantle, whereas the lithosphere is moving relatively west, or along sinusoidal flow lines based on tectonic features (Doglioni, 1990;
Doglioni et al., 1999), and confirmed by present-day space geodetic data (Hefflin et al., 2003). This flow coincides quite well with the fast direction detected for mantle seismic anisotropy in the asthenosphere (Montagner, 2002; Gaboret et al., 2003), which supports the model of relative flow in the decoupling zone (Doglioni et al., 2003). Because the ridge moves west at the sum of the speeds of the Pacific and Nazca plates divided by 2, the underlying asthenospheric melting area (where the mantle is undergoing depletion) is also moving west (Fig. 8). The viscosity of the asthenosphere below the Nazca plate should then be higher than beneath the Pacific plate; therefore, the Nazca plate and the subasthenospheric mantle should be better coupled (i.e., the Nazca plate moves west slower; Fig. 8). This model predicts that new fertile asthenospheric mantle is always available along the ridge, and deep continuous refeeding of the source for generating new oceanic crust is not needed. Kinematically it may be compared with a mantle bubble that shifts west along the area of plate separation.

Mantle tomography of the Pacific (Shapiro and Ritzwoller, 2002) shows higher velocities at ~200–250 km beneath Hawaii (in the middle asthenosphere), whereas velocities are relatively lower in shallower layers, supporting the idea of a deep “harder” asthenosphere at the base of the area from which the magmatism rises. Analysis of shear-wave splitting in the Pacific region shows a pattern in the asthenosphere that is parallel to the absolute motion of the Pacific. However, as shown by the Hawaiian chain, this pattern is disturbed by irregularities, testifying to some anisotropy in mantle flow (Montagner, 2002). This observation supports the presence of some local asperities (e.g., an increase in viscosity where higher seismic velocities are reported), which could trigger local shear heating in the asthenosphere. The proposed asthenospheric anisotropy in the model presented here may be tested in the future by more refined seismic tomography studies.

A global sinusoidal eastward motion of the subasthenospheric mantle relative to the lithosphere can also be inferred by the asymmetry of oceanic ridges that are 100–300 m shallower on their eastern flanks. An eastward-shifting asthenospheric mantle below the ridges, depleted and therefore lighter, can account for this asymmetry (Doglioni et al., 2003). The asymmetry of subduction zones cannot be ascribed merely to the age of the downgoing oceanic lithosphere, as it also persists where continental lithosphere subducts (Doglioni et al., 1999). Flow in the asthenospheric mantle could also explain the horizontal asymmetry of seismic anisotropy detected worldwide by Gung et al. (2003).

CONCLUSIONS

The preferred location of hotspots in a shearing asthenosphere would seem to discount the existence of hotspot reference
frames that include both intraplate and ridge-centered hotspots. A hotspot reference frame using only Pacific hotspots seems to be more reliable due to their relative fixity. However, in every hotspot reference frame, the Pacific plate is the fastest-moving plate in the world, and it also contains the most diffuse intraplate magmatism, regardless of its margins. Fastest velocity implies largest potential shear heating at its base. Classic, deep lower mantle sources (Morgan, 1971) for Pacific intraplate hotspots imply ~103–118 mm/yr shearing between the lithosphere and subasthenospheric mantle, as indicated by the age progression of the volcanic tracks. In this view, hotspots are generated either by fluid jets or anomalously high temperatures in the deep mantle (e.g., Schubert et al., 2001).

Alternatively, an asthenospheric mantle source for the same hotspots is supported by petrological, geochemical, and kinematic data. A model in which the source of the magma is the asthenosphere (e.g., Smith, 1993; Anderson, 2000) allows two important deductions: (1) Shear heating generated in the asthenosphere by the motion of the overlying Pacific plate is large enough to generate the scattered and diffuse intraplate magmatism observed (Figs. 3 and 4), and (2) the velocity recorded in volcanic chains is a function of the depth to the source of the magmatism in the asthenosphere (Fig. 1) and its trend relative to plate motion direction (Figs. 5 and 6). The shear between the lithosphere and subasthenospheric mantle is expected to be greater than 103–118 mm/yr, because of the missing shear in the asthenosphere below the hotspot source. This estimate provides a global net rotation of the lithosphere toward the west that is larger than the present estimate of 49 mm/yr (Gripp and Gordon, 2002). In this interpretation, all plates on Earth move westward relative to the subasthenospheric mantle, although at different velocities, and with a sinusoidal trend (Doglioni, 1993). Therefore, in contrast with the predictions of the classic hotspot reference frame (Fig. 7, option 1), the subasthenospheric mantle beneath the Nazca plate would move eastward relative to the lithosphere (Fig. 7, option 2), thus providing a mechanism for continually providing new fertile asthenosphere to melt beneath the westward-migrating EPR (Fig. 8). With these velocities relative to the subasthenospheric mantle, the net rotation of the entire lithosphere at its “equator” could be faster than 90 mm/yr.

Most of the Pacific asthenosphere has a viscosity around $5 \times 10^{17}$ Pa·s (Pollitz et al., 1998). However, lateral variations in geochemistry and fluid depletion can result in local increases in viscosity. Greater shear between the lithosphere and subasthenospheric mantle could generate excess heating exceeding 100 °C if the asthenosphere contains local anomalies in which the viscosity is higher ($4 \times 10^{19}$ to $10^{20}$ Pa·s). For the maximum temperature increases, extra melting can be generated, which may be responsible for scattered superficial magmatism. An increase of ~100 °C would shift the adiabat curve of uprising mantle to the right (Fig. 4), shifting the asthenospheric mantle to the field of more extensive melting (Green, 2003). This model re-creates the work of Shaw and Jackson (1973) and can explain Pacific hotspots often attributed to deep mantle plumes as stress-generated intraplate melt anomalies in the asthenosphere. The model also explains the observed periodicity and variability in magma segregation, because strong shear heating decreases viscosity and therefore also inhibits melting until a higher viscosity is regained and shear heating recommences. The origin of such hotspots would therefore be different from those persistently located on oceanic ridges and sourced in the asthenosphere, where more abundant water in the mantle (wetspots) more strongly affects the extent of melting (Bonatti, 1990; Asimov and Langmuir, 2003). The main differences between hotspots and wetspots (e.g., Hawaii versus Ascension) may be summarized in terms of 3D kinematics and petrology:

1. **Hotspots.** The plate moves relative to a deeper source and the volcanic trail is independent of any plate margin. The age of the trail is much younger than the host lithosphere. The source of the hotspot is in or below the asthenospheric décollement. The melting temperature is higher (1400–1500 °C, possibly caused by shear heating).

2. **Wetspots.** The wetspot moves with the oceanic ridge, and has an age close to that of the host oceanic crust (i.e., a speed close to that of the plate boundary). The source of the wetspot is above the asthenospheric décollement. The melting temperature and pressure are lower (perhaps <1300 °C, possibly caused by more water in the mantle).

Because wetspots are sourced above the plate-tectonic décollement and do not move with the mantle but rather migrate at the same velocity as the oceanic ridge, they exhibit different surface kinematics. For example northeast-trending, water-rich, parallel anomalies in the asthenospheric mantle generate surface volcanic trails with the same northeast-trending orientation but propagating to the southwest (e.g., Walvis and Cameroon ridges, Fig. 9). Similar but northwest-trending and southeast-propagating volcanic trails can be generated in the paired plate to the west (e.g., Rio Grande Rise). The northwest-trending Emperor hotspot track can be generated by a north-northeast-trending asthenospheric anomaly located in the décollement surface, and the resulting magmatism propagates toward the southwest. In both these cases, there is an angle between absolute plate motion and mantle structural anisotropy, but the resulting volcanic tracks trend in the opposite directions (compare Figs. 5 and 9). In the case of the Hawaiian track, the directions of plate motion and mantle anisotropy seem to coincide.

There exist traces that extend from intraplate to ridge-centered locations and vice versa (e.g., Foundation, Nazca–Easter-Tuamotu, Cobb), plus ridge-crossing traces (e.g., New England–Great Meteor, Réunion, Kerguelen). Is each of those tracks really linked to the same source (e.g., is Réunion really connected to the Chagos-Laccadive system), or are they two independent magmatic systems? Are some of these tracks related to transtensional tectonics along transform zones? Could shear heat-
ing and wet melting occur along the same trail? Can shear heating persist beneath a ridge, or is there a switch to a water-driven process? These are questions that remain to be answered.

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Figure 9. Hypothetical reconstruction of the south-Atlantic–type migrating volcanic ridges. An anomalously water-rich asthenospheric mantle, or wetspot (sensu Bonatti, 1990), oriented oblique to absolute African plate motion and the mid-Atlantic Ridge (MAR) migration could generate a southwest-oriented rejuvenated volcanic track. A similar, mirrorlike track (northwest trending, southeast propagating) could form in the South America plate. This model could explain why the age progression of the volcanic trail is oblique to the transform faults.


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