The Thermal State of the Upper Mantle;

No Role for Mantle Plumes

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Abstract

A variety of geophysical data indicates that long wavelength temperature variations of the asthenosphere depart from the mean by ± 200°C, not the ± 20°C adopted by plume theoreticians. The 'normal' variation, caused by plate tectonic processes, such as subduction cooling, continental insulation and small-scale convection, encompasses the temperature excesses that have been attributed to hot jets and thermal plumes. Geophysical estimates of the average potential temperature of the upper mantle are about 1400°C. Asthenospheric convection at ridges, rifts and fracture zones and at the onset of continental breakup is intrinsically 3D, giving rise to shallow pseudo-plume-like structures without deep thermal instabilities. Deep narrow thermal plumes are unnecessary and are precluded by uplift and subsidence data, and widespread and repeated volcanism associated with many regions of excess magmatism. The locations and volumes of 'midplate' volcanism appear to be controlled by lithospheric architecture, stress and cracks, and small-scale convection.

Introduction

More than 25 years ago Verhoogen (1973), Elder (1976) and Richter (1973) pointed out the importance of lateral temperature gradients and small scale convection in the upper mantle in problems of magma formation. More recently, theoretical models of terrestrial magmatism assume that the normal state of asthenospheric mantle is isothermal, and thus static, and also subsolidus. This applies to both passive (e.g. Boutilier and Keen, 1999) and plume (e.g. White and McKenzie, 1989) type models. However, systems cooled from above or having lateral temperature, conductivity or radioactivity gradients at the top will develop small-scale convection. This topside driven convection can be an order of magnitude faster than plate rates or rifting rates (Korenaga, 2000) and therefore cannot be ignored or treated as a small perturbation. Rapid vertical convection increases the melt delivery to the surface in regions of extension without an increase in mantle temperature. In theories of terrestrial magmatism, crustal thickness is often used as a proxy for mantle temperature. The plume hypothesis, to a large extent, is based on the hypothesis that the 'normal' state of the mantle is isothermal, cold, subsolidus, static, refractory, dry and homogeneous, and that 3D, focusing and small-scale convection effects are not important. The locations of volcanoes and the variations in crustal thickness and extrusion rates, however, depend on more than temperature. The ability of rift-induced dynamic convection to explain large igneous provinces from 'normal' temperature mantle obviates the need for hot deep mantle plumes (King and Anderson, 1998; Boutilier and Keen, 1999). Topside tectonics also explains plate motions and continental breakup (Elder, 1976; Lowman and Jarvis, 1999; Richter, 1973).

McKenzie and Bickle (1988) assumed the upper mantle to be homogeneous and more-or-less isothermal. They adopted a 'cold' subsolidus potential temperature of 1280°C ± 20°C and assumed that temperatures that are 150° to 200°C higher than this are due to localized hot jets. This type of isothermal mantle implies either no convection, or very high (> 10^{12}) Rayleigh number convection (e.g., Niemela et al., 2000).

Geophysical estimates of upper mantle temperatures are more than 100°C hotter (e.g. Anderson and Bass, 1984; Anderson, 1989; Kaula, 1983; Hofmeister, 1999), and the normal variability is about 10 times greater than adopted by McKenzie and Bickle (1988) and subsequent advocates of the thermal plume hypothesis (e.g. White, 1988). There is very little support for the cold isothermal static asthenosphere hypothesis, a corollary of the plume hypothesis. To a large extent the plume hypothesis is based on this strawman model. If normal mantle temperatures are 1400° ± 200°C, or even 1350° ± 150°C, there is no thermal requirement for hot mantle plumes. Small scale convection associated with ridges, rifts and edges is intrinsically 3D, giving rise to concentrated, but shallow, plume-like upwellings (Richter, 1973; Parmentier and Phipps Morgan, 1990; Shen and Forsyth, 1992). Thus, there is also no geometric requirement for the deep mantle plume hypothesis. Geophysical and petrological estimates of mantle potential temperatures (the P = 0 extension of the adiabat) are summarized in this paper. A shallow source for geochemical variations was addressed earlier (Anderson, 1994, 1995).

Temperature Variations in the Upper Mantle

The range of temperatures in the upper mantle, below the thermal boundary layer (‘fully convective region’ in the nomenclature of Kaula (1983)) is easier to constrain than the absolute temperature. Bathymetry, rate of seafloor subsidence, heat flow, and depths of the 410 and 650 km discontinuities are all functions of temperature and complement the standard petrological approaches utilizing crustal chemistry and thickness variations. There is a remarkable consistency in estimates derived from these various datasets (Table 1).

Kaula (1983) estimated the minimal upper mantle temperature variations that are consistent with observed heat flow and plate velocities. At the fully convective level, about 280 km depth, temperature variations are at least ± 180°C, averaged over 500 km spatial dimensions. This is in contrast to the assumption of McKenzie and Bickle (1988) that at this depth all geotherms are horizontal.

Temperature variations along the global midocean ridge system, based on petrology and crustal thickness are about 200°C (Klein and Langmuir, 1987; Kane and Hayes, 1994). This represents about half the global range (Kaula, 1983) and agrees with estimates based on heat-flow and plate motions. Along-ridge variations in bathymetry and subsidence rates also imply upper mantle temperature variations of 200°C (Perrot et al., 1998), even when deep trench areas are avoided.

The composition of near-ridge peridotites has also been used to infer a temperature range of 200°C along the ridge system (Bonatti, 1990) and to infer that some so-called hotspots are actually wet-spots of normal temperature. In particular the Azores platform appears to be related to near-ridge fracture zones and transform faults and is not underlain by hot mantle (Azevedo and Portugal, 1999).

The average thickness of the transition region (400 to 650 km depth) constrains the mantle temperature and the variation in thickness constrains the temperature variation (Anderson, 1967; 1989). Flanagan and Shearer (1998) obtain 244 ± 32 km as the thickness of the transition region. This gives a temperature variation of ± 120°C to ± 230°C depending on choice of thermochemical parameters (Anderson, 1989; Agee, 1998). Locally, in the vicinity of deep slabs, the transition region may get as thick as 290 km, (e.g., Clarke et al., 1995). This implies local temperatures about 100˚C colder than the long wavelength global extreme. A recent detailed study gives 250 ± 10 km as the global range (Chevrot et al., 1999) which gives a $\Delta T \leq \pm 100°C$ in the transition region, consistent with inferences of Li et al. (1998). If these values are correct, then ΔT may decrease with depth. Melbourne and Helmberger (2000) determined that transition zone thicknesses under the entire East Pacific Rise (EPR) are indistinguishable from the global mean (PREM; Dziewonski and Anderson (1981)) and under the Canadian shield. The inferred temperature variations below 400 km are less than 25˚C. This region contains three or four proposed hotspots yet they do not influence transition zone temperatures, suggesting shallow roots, not only for ridges, but also for regions of excess volcanism. Variations inferred for the lower mantle, above D″, are also low (Duffy and Ahrens, 1992; Gong et al., 2000), less than ± 50°C. This decrease with depth may be due to removal of the colder upper parts of the slab at depths above 400 km. Continental insulation (Anderson, 1982) and small-scale convection may also primarily affect the shallow mantle.

The Chevrot et al. (1999) study shows no transition zone thinning under Iceland, Hawaii, Easter, Afar, Yellowstone or Cameroons, all considered, by some, to be hot plumes. The shallow mantle in some of these regions, and also the EPR, has low seismic velocities, at long wavelengths; high temperatures and large temperature fluctuations apparently do not extend to 650 km. Regional studies near subduction zones imply transition zone temperatures 200-300°C colder than average (Clarke et al., 1995).

The temperature increase at the base of the mantle is estimated to be between 1000° - 2000°C (Williams, 1998), much greater than excess temperature near ‘hotspots’ and long-wavelength lateral temperature changes in the upper mantle. This strongly indicates that temperature variations in the upper mantle have nothing to do with a TBL at the base of the mantle, or that the mantle is chemically layered. The lateral variations in temperature at the top of the mantle, and those inferred in the absence of plumes and lower thermal boundary layer participation are more than adequate to explain terrestrial magmatism and its variety.

It can be noted that a temperature rise of 200°C can bring an upper mantle rock from subsolidus to one that is 20% molten. Although the deepest and coldest parts of the global ridge system are about 175°C colder than average and are melt starved, they still provide some basalts (Bonatti et al., 1993, 1994; Christie et al., 1998; Lanyon et al., 1995). This suggests that "average" mantle, or at least "average ridge" mantle, is above the solidus, even before adiabatic decompression.

The above estimates of temperature fluctuations are consistent with the ± 10% variability which accompanies "normal" convection (Elder, 1976; Lowman and Gable, 1999; Niemela et al., 2000) and are of the order required to drive 3D shallow mantle small scale convection and plume-like instabilities (Davaille and Jaupart, 1994). These shallow plume-like instabilities can deliver even larger volumes of melt from normal temperature mantle than 2D rolls (Korenaga, 2000) or hot deep mantle plumes (Cordery et al., 1997).

Absolute Temperature

The absolute temperature of the mantle is harder to constrain than the temperature variation, but values estimated from mineral physics, seismology, geodynamics, heat flow and petrology are consistent (Kaula, 1983; Anderson and Bass, 1984; Duffy and Anderson, 1989; Anderson, 1989). These approaches give a mean potential temperature of about 1400°C for the upper mantle. Equation of state fits to the lower mantle yield potential temperatures of 1500°C

(Zhao and Anderson, 1994; Stacey, 1992) to 1730°C (Stixrude et al., 1992). The average conductive geotherm (Hofmeister, 1999) intersects the upper mantle adiabat slightly below 80 km depth, and the wet peridotite or eclogite solidus at shallower depth. In the warmer parts of the mantle the geotherm may be on or above the solidus to depths as great as 300 km (Anderson and Bass, 1984; Anderson, 1989). These studies also indicate that colder parts of the mantle, e.g. sub-shield, are below the solidus throughout.

The mean temperature under the global spreading ridge system is slightly more than 1500°C and under subduction zones is about 1200°C, at a depth of 280 km, averaged over lateral dimensions of about 500 km (Kaula, 1983). The comparable potential temperatures are about 1410° and 1110°C. The coldest part of the upper mantle should be just above cold subducting slabs. Even here the inferred temperature from petrology (Sisson and Brunto, 1998; Tatsumi, 1994) is well above the McKenzie-Bickle average temperature.

Global tomographic studies show that both ridges and hotspots occur preferentially over broad regions of hotter than average mantle (Anderson et al., 1992; Wen and Anderson, 1995, 1997a). If the average temperature of the mantle is close to the melting point, (Anderson and Sammis, 1970; Anderson, 1989) the inference is that the ridge and hotspot mantle, in fact, most of the oceanic mantle, is near the solidus to depths greater than 250 km, consistent with seismic inferences (Anderson and Bass, 1984).

Geophysical modeling of the oceanic heat flow and bathymetry imply a temperature increase across the oceanic thermal boundary layer of 1400°C (e.g. Hofmeister, 1999; Stein and Stein, 1992). This assumes isotropic thermal conductivity, a chemically homogeneous boundary layer, and no shear heating. When these effects are taken into account, the temperature can increase by more than 200°C unless buffered by melting (Hearn et al, 1997). These results imply that the potential temperature of the oceanic asthenosphere is at least 1400°C. The rarity of magmas with such high temperatures suggest that they are too dense to be eruptable except in exceptional or transient situations. Seismic velocities in different tectonic provinces converge near 400 km depth, but at 350 km the inferred temperatures under cratons are at least 100°C colder than under ridges, assuming similar compositions and phases (Anderson and Bass, 1984). The temperatures which are consistent with recent phase equilibria data for the olivine-spinel phase change are 1410° to 1760°C at 400 km depth (Morishima et al., 1994). An estimate of the maximum potential temperature of 'normal' mantle below 400 km can be obtained from the minimum long-wavelength thickness of the transition region (~210 km; Flanagan and Shearer, 1998). This gives 1440°C, using thermochemical calculations of Agee (1998).

Extrapolations of the lower mantle adiabat are generally greater than estimates of the upper mantle potential temperature (Hofmeister, 1999), suggesting stratified mantle convection. The chemical boundary may be near 1000 km depth (Wen and Anderson, 1995, 1997b), the top of Bullen's region D.

In summary, the average potential temperature of the upper mantle appears to be about 1400˚C with an uncertainty of ± 50˚C. The variability of upper mantle temperatures is about ± 200˚C. The potential temperature of the lower mantle appears to be at least 100˚C hotter than the upper mantle, and the long wavelength variability in transition zone and lower mantle temperatures is about ± 100˚C. The normal expected range of ΔT, caused by convection and plate tectonic processes, without plumes, is about ± 200˚C (Anderson, 1998a; Lowman and Jarvis, 1999; Elder, 1976). This is much larger than assumed by plume theoreticians but much lower than expected if temperature excesses are imported from the core-mantle boundary.

Convection driven by edge effects and variable lithospheric thickness can deliver melts at the volumes and rates required at large igneous provinces from mantle with temperatures in the 'normal' range (King and Anderson, 1998; Boutilier and Keen, 1999; Korenaga, 2000). Shallow plume-like 3D effects are even more effective (Korenaga, 2000), and likewise do not require substantial heating from below.

Hotspot Temperatures

Plumes are a hypothetical form of convection based on the premise that excess magmatism require localized regions of high temperature. Since normal convection and plate tectonic processes can cause variations of ± 200˚C and excess magmatism can also be caused by small-scale convection and magma focusing, there is no *à priori* need for deep mantle plumes. Convection along spreading ridges and continental boundaries is intrinsically 3D and gives pseudo-plume structures that concentrate magmatism without deep boundary layer instabilities (Parmentier and Phipps Morgan, 1990; Richter, 1973). The association of large igneous provinces with triple junctions, continental and craton margins, ridges and fracture zones suggests lithospheric control rather than a unique form of convection controlled by the core-mantle boundary (Anderson, 1998b). Herein I summarize geophysical estimates of hotspot temperatures and conclude that these are in the range of "normal" mantle temperatures.

Ribe et al. (1995) and Feighner et al. (1995) show that hotspots are much colder than once thought (e.g. Schilling, 1991). They derive temperature 'excesses' of 57˚C, 51˚C and <70˚C for the Azores, Galápagos, and Iceland. Ito and Lin (1995b) used bathymetry and gravity to estimate near-ridge temperatures adjacent to hotspots. The temperature excesses are generally between 50˚ and 150˚C (Iceland, Azores, Tristan and Easter). Interestingly, temperatures of

Galápagos basalts (1186˚C ± 30˚C) are less than along the nearby ridge segment (Fisk et al., 1982), and the basalts near the center of the conjectured plume are more depleted (MORBish) than those away from the center (Geist et al., 1988).

Schilling (1991) had earlier attempted to infer excess temperatures of plumes from bathymetry. Values for 13 proposed hotspots fall in the narrow range of + 162°C (for Tristan) to + 278°C (for Circe), much higher than later estimates by Ito and Lin (1995b) and others. There is no relation to hotspot-ridge distance or discharge rate, as expected from plume theory. These are upper bound temperature estimates since density changes due to partial melting and restite layers are ignored. Schilling quotes an uncertainty of 50% in his estimates. Therefore, even Schilling's estimates fall within the temperature variations of normal upper mantle processes (convection, plate tectonics).

Local variations in bathymetry and subsidence rates of oceanic lithosphere imply temperature variations of 100° to 200°C in regions picked to avoid 'hotspot influence' (Kane and Hayes, 1994). These are superposed on interocean differences of 20° to 35°C. Thus, the temperature 'excesses' attributed to ridge-hotspot interactions are within the range of normal mantle temperature variations. The much larger excess temperatures required by the plume hypothesis (e.g. Cordery et al., 1997) are not supported by the data. Iceland, Azores, Tristan, Galápagos and Easter are the five hotspots that impose the most prominent bathymetric and geochemical anomalies, yet these imply temperature anomalies of less than 150°C (Ito and Lin, 1995b). They can be regarded as near-ridge fracture zone and edge phenomena and regions of small-scale convection and lithospheric extension (Sykes, 1978; Richter, 1973; Favela and Anderson, 1999).

Mantle temperatures can also be inferred from guyot heights (Caplan-Auerbach et al., 2000). Most seamounts and volcanic islands were emplaced on seafloor overlying mantle less than 100°C hotter than average. Many intraplate volcanic regions have no thermal anomaly at all (Pratt-Welker seamounts, Midpacific Mountanins, Trinidade, Japanese seamounts). The median thermal anomaly for the Easter and Emperor chains is 150°C. Guyots along the Hawaiian, Marquesas, Louisville, Marshall and Darwin Rise imply temperature excesses of less than 100-200°C.

Some 'hotspots' are cold, and some are wet. Takahashi et al. (1998) argues that ocean island basalts are from fertile mantle (e.g. piclogite) rather than hot mantle, and that most estimates and assumptions of ocean island and continental flood basalt temperatures and volumes are too high. They calculate that magmas attributed to hotspots originate in mantle of 1400°C or less.

Korenaga (2000) determined that the mantle temperature associated with the breakup of Greenland from Europe and the North Atlantic igneous province was 1270˚ - 1350˚C throughout the rifting process and that changes in the volumes of extrusives were related to small-scale convection. The crust of the Kolbeinsey Ridge, just north of Iceland, implies a constant temperature of 1320-1360˚C for the past 22 Ma (Kodaira et al., 1998). This hotspot province is well within the range of 'normal' mantle temperatures. Tegner et al. (1998) assemble other arguments against the plume hypothesis in this area.

Theoretical or predicted plume temperatures (> 1600°C) are much higher than petrological and geophysical estimates (Campbell and Griffiths, 1990; Richards et al., 1989; Cordery et al., 1997; Tegner et al., 1998). Seismic velocities indicate that the Iceland crust and shallow mantle is cold, probably colder than the East Pacific Rise. Partial melt may exist beneath Iceland, but if so, it occurs 25 km below Moho, much deeper than under spreading ridges (Menke et al., 1998). The viscosity of the crust and upper mantle under Iceland also indicate relatively cold temperatures (Pollitz and Sacks, 1996). The temporal evolution of Greenland and Iceland basalts is inconsistent with plume models and call for important lithospheric control.

Seismic investigations of Hawaii indicate a deep cold root extending to at least 100 km (Woods and Okal, 1996; Priestley and Tilman, 1999). The Woods and Okal (1996) study shows that the upper 200 km under Hawaii is similar to normal Pacific mantle. Inferred temperatures are relatively low (1463°C), and the top of the melt zone is deeper than in the plume calculations of Watson and McKenzie (1991). There is no evidence from seismology that the transition zone under Hawaii is thin (hot) (Chevrot et al., 1999). Multiple ScS phases, bouncing under Hawaii, and the Galápagos, show normal, or fast, mantle velocities (Best et al., 1974; Sipkin and Jordan, 1976). Some factor other than temperature, such as lithospheric architecture and stress, is controlling the locations and volumes of midplate volcanism (Anderson, 1998a, b; Favela and Anderson, 1999; Tegner et al., 1998). Most volcanoes of all kinds occur above upper mantle that is seismically slow, or hot, but specific locations appear to be controlled by the lithosphere.

Causes of Lateral Temperature Gradients

Temperature variations at the top of the convecting mantle are caused by slab cooling, cratonic roots and continental insulation (Anderson, 1998a). These are imposed lateral temperature changes in contrast to the accidental or induced gradients set up by Rayleigh Bénard convection driven by bottom heating and vertical temperature gradients. The imposed variations amount to ± 200˚C and have various characteristic length scales. The shears imposed by moving plates and rise of the asthenosphere between cratons impose other length scales and generate another dimension of motion (horizontal rolls and vertical stalks) into what appear to be 2D problems (linear ridges and rifts). Lateral temperature gradients alone can induce 3D upwellings. The plume hypothesis ignores these effects and the resultant shallow plume-like upwellings and attributes all such features to deep thermals. Strictly 2D convection is confined to low viscosity and nearly isothermal situations with uniform boundaries.

Small-Scale Convection

Malamud and Turcotte (1999) calculate that 5242 plumes are required to satisfy terrestrial heat flow observations, or one every 156 km at the core-mantle boundary. Small-scale convection is an alternative, but they rule this out because “the concept of plume(s)…is now widely accepted,” and the required asthenospheric viscosity, about 10^{18} P_a.s is lower “than most estimates…”. Actually, such viscosities are appropriate for the asthenosphere (Cathles, 1975; Richter and McKenzie, 1978; Korenaga, 2000; Hirth and Kohlstedt, 1996). Sublithospheric mantle flow is generated by lithospheric architecture and stress, and melt instabilities (e.g., Schmeling and Marquart, 1993).

Summary

The geophysical data that constrains the lateral variations of temperature below the plate include: bathymetry, subsidence rates, heat flow, global plate motion modeling, depths to mantle phase changes, seismic velocities, thickness of the transition region and crustal thickness. These data imply temperature variations of ± 150° to ± 200°C, even when filtered to avoid ‘hotspot influence’ and subduction zones. There is good agreement between various geophysical estimates of ‘normal’ upper mantle temperature variations.

The potential temperature of the upper mantle is 1400° ± 200°C based on long-wavelength bathymetry, subsidence, heat flow, tomography, plate motions, discontinuity depths and petrology. The mean is more than 100°C hotter than assumed for 'normal' mantle in the plume hypothesis. Estimates of mantle temperature in the vicinity of 'hotspots' fall within this range. 'Hotspots' do not require 'excess' temperatures or hot plumes. Excessive magmatism and locations of volcanoes appear to be controlled by stress, lithospheric architecture, rift and edge-induced convection and focusing, not narrow hot jets from near the core (King and Anderson, 1998; Favela and Anderson, 1999). The very large variations in temperature that characterize the uppermost and lowermost 300 km of the mantle apparently are not transmitted to the transition zone and lower mantle. This is a new constraint on mantle dynamics.

The absence of appreciable thermal anomalies associated with hotspots and continental flood basalts (Czamanske et al., 1998; Korenaga, 2000) suggests that rapid fluxing of the asthenosphere through the melting zone (Boutilier and Keen, 1999; King and Anderson, 1998; Anderson, 1994; 1995) and 3D effects (Richter, 1973; Parmentier and Phipps Morgan, 1990) are responsible for excess magmatism, not hot mantle plumes. "Topside tectonics," is now a more mature and self-consistent theory (less contradictions, paradoxes and coincidences) than is plume theory and does not require an *ad hoc* initial singularity (e.g. Cordery et al., 1997) to get it started.

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TABLE 1. Long Wavelength Temperature Variations in the Sublithospheric Mantle

GLOBAL	
Heat flow, plate velocities (Kaula, 1983)	± 180ºC
Ridges; petrology (Klein and Langmuir, 1987)	± 125ºC
Ridges; subsidence (Calcagno and Cazenave, 1993)	± 100ºC
Transition Zone thickness	± 100ºC
Lower Mantle; seismic velocity (Yoneda and Spetzler, 1994)	± 112ºC
North America (Butler, 1984)	± 145ºC
Theoretical (Anderson, 1998)	± 200ºC
COLD REGIONS	
Cratons; Depth of phase changes (Li et al., 1998)	< -150ºC
Subduction (Anderson, 1997)	-150ºC
Deep ridges (Bonatti et al., 1994; Lanyon et al., 1995)	-150ºC*
Non-ridges (Kaula, 1983)	-100ºC to -250ºC
HOT REGIONS	
Supercontinent insulation/isolation (Anderson, 1998)	+200ºC
Ridges (Kaula, 1983)	+100º to +250ºC
Iceland (Sato and Sacks, 1989)	~ +120ºC
Iceland (Ribe et al., 1995)	< +70ºC
"Hotspots" (Ribe et al., 1995)	+50º to +70ºC
(White and McKenzie, 1989)	+150º to +200ºC
(Skogseid et al., 1992)	+50º to +130ºC
(Ito and Lin, 1995a)	+50º to +150ºC
(Schilling, 1991)	+162º to +278ºC

* from ridge mean

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