

Excess temperature of mantle plumes: The role of chemical stratification across D''

C. G. Farnetani

Institut de Physique du Globe de Paris, France

Abstract. The temperature increase across the thermal boundary layer at the base of the Earth's mantle (D'') may be as high as 1000-1300°C, while the excess temperature of mantle plumes, inferred from petrological studies, is ca. 200-300°C. These two estimates are generally considered separately, although this should not be the case if plumes do originate from D''. This paper investigates the evolution of plume excess temperature from the source region in the deep mantle to lithospheric depths. A finite element model in axisymmetric geometry is used to explore the effects of a number of model variables. The results show that if the source region is *only* a thermal boundary layer, the plume excess temperatures are unrealistically high, since entrainment of surrounding mantle does not cool significantly a large plume head rising through the mantle. I then investigate the role of a chemically denser layer at the base of D'', considering a range of values for its thickness and excess density. The presence of a 30 km thick layer, 5% chemically denser than the surrounding mantle, buffers the excess temperature of plumes to plausible values. The results suggest that chemical stratification across D'' may govern the excess temperature of mantle plumes.

Introduction

The D'' region at the base of the Earth's mantle is an important thermal boundary layer and it is the most likely source region of mantle plumes [e.g., *Stacy and Loper, 1983*]. Another thermal boundary layer could exist between the upper and the lower mantle, as predicted by models of layered mantle convection [*Richter and McKenzie, 1981*]. However, an ample range of geophysical observations weighs against the transition zone being a complete barrier to mantle flow [e.g., *Davies and Richards, 1992*].

Jeanloz and Morris, [1986] estimated a temperature increment of 800-1000°C across D'', based on the mismatch between calculated adiabatic temperatures in the lower mantle and in the core. A temperature increment in excess of 1300°K is inferred from recent melting experiments of iron and iron alloys at core pressures [*Boehler et al., 1995*]. The excess temperature across D'' is considerably higher than the estimated temperature anomalies of mantle plumes at lithospheric depths. Plume excess temperatures of ca. 250°C are inferred from the petrology and isotope geochemistry of hotspot lavas [e.g., *Schilling, 1991*]. Numerical models for Hawaii indicate excess temperatures of ca. 280°C [*Watson and McKenzie, 1991*], while dynamical models of plume-

lithosphere interaction show that plume head excess temperatures of 350°C are sufficient to generate melt volumes comparable with those of continental flood basalts, even for the restricted case of thick, not-rifting lithosphere [*Farnetani and Richards, 1994*].

In the last decade fluid dynamical laboratory experiments have shown that thermal plumes entrain a large volume fraction of surrounding fluid as they ascend [*Griffiths and Campbell, 1990*]. Entrainment of colder mantle would therefore reduce the excess temperature of a mantle plume. *Farnetani and Richards [1995]* question such an important role of entrainment. In their numerical models a plume generated by a thermal boundary layer instability entrains only a small fraction of surrounding mantle. The low degree of entrainment (10-20%) quantified within the region of the plume head undergoing partial melting suggests that entrainment is not an efficient mechanism to control plume excess temperatures.

The interpretation of D'' as a chemically distinct layer is still a matter of debate, however there is a growing consensus on the existence of chemical heterogeneities within D'' [e.g., review by *Loper and Lay, 1995*]. Seismological observations and mineral physics experiments are compatible with the existence of two types of metal-bearing boundary layers [*Manga and Jeanloz, 1996*]: (i) a layer 0-40 km thick, rich in Fe-alloy, whose estimated seismic velocities may be comparable to the ultralow-velocity observed beneath the central Pacific by *Garnero and Helmerger [1996]*, (ii) a layer 0-300 km thick, rich in FeO and FeSi, whose seismic velocities may be comparable to the thick low-velocity layers detected by *Vinnik et al., [1995]*.

In this paper I focus on the evolution of plume excess temperature from the source region in the deep mantle to lithospheric depths. Two main questions are addressed: Which processes control the excess temperature of mantle plumes? Can we reconcile the inferred high temperature contrast across D'' and the estimated excess temperatures of mantle plumes?

I use a finite element model to solve for the dynamics of solid state convection and for the advection of a chemical field. The chemical field allows me to investigate the role of chemically denser material initially confined at the base of the thermal boundary layer. For most model the chemical boundary layer is 30 km thick, and it may be up to 10% denser than the surrounding mantle. My results indicate that: 1) Entrainment does not govern the excess temperatures of plumes. For models without chemical stratification the plume head excess potential temperature in the uppermost mantle is only a few hundred degrees lower than the boundary layer excess temperature. 2) The presence of chemical stratification across the source region has a strong effect in buffering the final excess potential temperature of plume heads. Although my models are a considerable simplification of the complex processes occurring within D'', they indicate that chemical

Copyright 1997 by the American Geophysical Union.

Paper number 97GL01548.
0094-8534/97/97GL-01548\$05.00

heterogeneity at the base of the mantle may control the excess temperature of the buoyant material feeding mantle plumes.

The Numerical Model

The equations of conservation of mass, momentum, and energy for solid-state flow in the mantle are solved in axisymmetric cylindrical geometry, using a finite element code described in detail elsewhere [Farnetani and Richards, 1995]. The code also solves for the advection equation of a continuous function which tracks the concentration of chemically distinct source material. The chemical field, thereafter called the C-field, is a dimensionless number denoting the concentration of denser material. Thermochemical convection models are notoriously difficult to calculate [vanKeken *et al.*, 1997] and the field method suffers from some numerical diffusion. I apply a corrector algorithm to reduce the spurious diffusion [see Farnetani and Richards, 1995], and the Lewis number (Le =thermal diffusivity/chemical diffusivity) is $Le=100$. The physical constants defining the Rayleigh number ($Ra=g\rho_o\alpha\Delta Td^3/\kappa\eta_m$) are: $g=10\text{ m s}^{-2}$, mantle density $\rho_o=4000\text{ kg m}^{-3}$, thermal expansion coefficient $\alpha=2\times 10^{-5}\text{ }^\circ\text{C}^{-1}$, $\Delta T=T_m-T_s$, ($T_m=1300^\circ\text{C}$ mantle potential temperature, $T_s=0^\circ\text{C}$ surface temperature), depth of the cylinder $d=2900\text{ km}$, thermal diffusivity $\kappa=10^{-6}$

$\text{m}^2\text{ s}^{-1}$, mantle viscosity $\eta_m=3\times 10^{21}\text{ Pa s}$. The density varies with temperature and composition as $\rho=\rho_o-\Delta\rho_t+\Delta\rho_{ch}$ where $\Delta\rho_t=\rho_o\alpha\Delta T$ is the thermal density deficiency and $\Delta\rho_{ch}=\rho_o\beta C$ is the chemical excess density. For $\beta=0$ no chemical buoyancy is associated to the chemical field, $\beta=0.03$ and $\beta=0.05$ correspond, respectively, to source material 3% and 5% chemically denser than the surrounding mantle. The viscosity depends only upon temperature, according to an exponential law $\eta(T)=\eta_m\exp(-b(T-1))$, T is the nondimensional potential temperature, b is a constant chosen to give a maximum dynamic viscosity contrast of 100 between the plume and the surrounding mantle. This viscosity ratio is modest compared to estimates for olivine rheology, but such a low ratio guarantees good resolution. The initial and boundary conditions are designed to model a narrow upwelling plume from a boundary layer at the bottom of the mantle. The constant temperature imposed at the bottom satisfies the modeled potential temperature contrast across the thermal boundary layer (ΔT_{bl}). Real temperatures (T_r) can be estimated from potential temperatures (i.e., the temperature that the material would have if brought adiabatically to the Earth's surface) using $T_r=T\exp(gz/C_p)$, $C_p=1200\text{ J kg}^{-1}\text{ }^\circ\text{K}^{-1}$ is the specific heat at constant pressure, z is depth. The reference model has $\Delta T_{bl}=800^\circ\text{C}$, which corresponds to a real excess temperature $\Delta T_{rbl}=1300^\circ\text{C}$ for $\alpha=2\times 10^{-5}\text{ }^\circ\text{C}^{-1}$ (or $\Delta T_{rbl}=1020^\circ\text{C}$

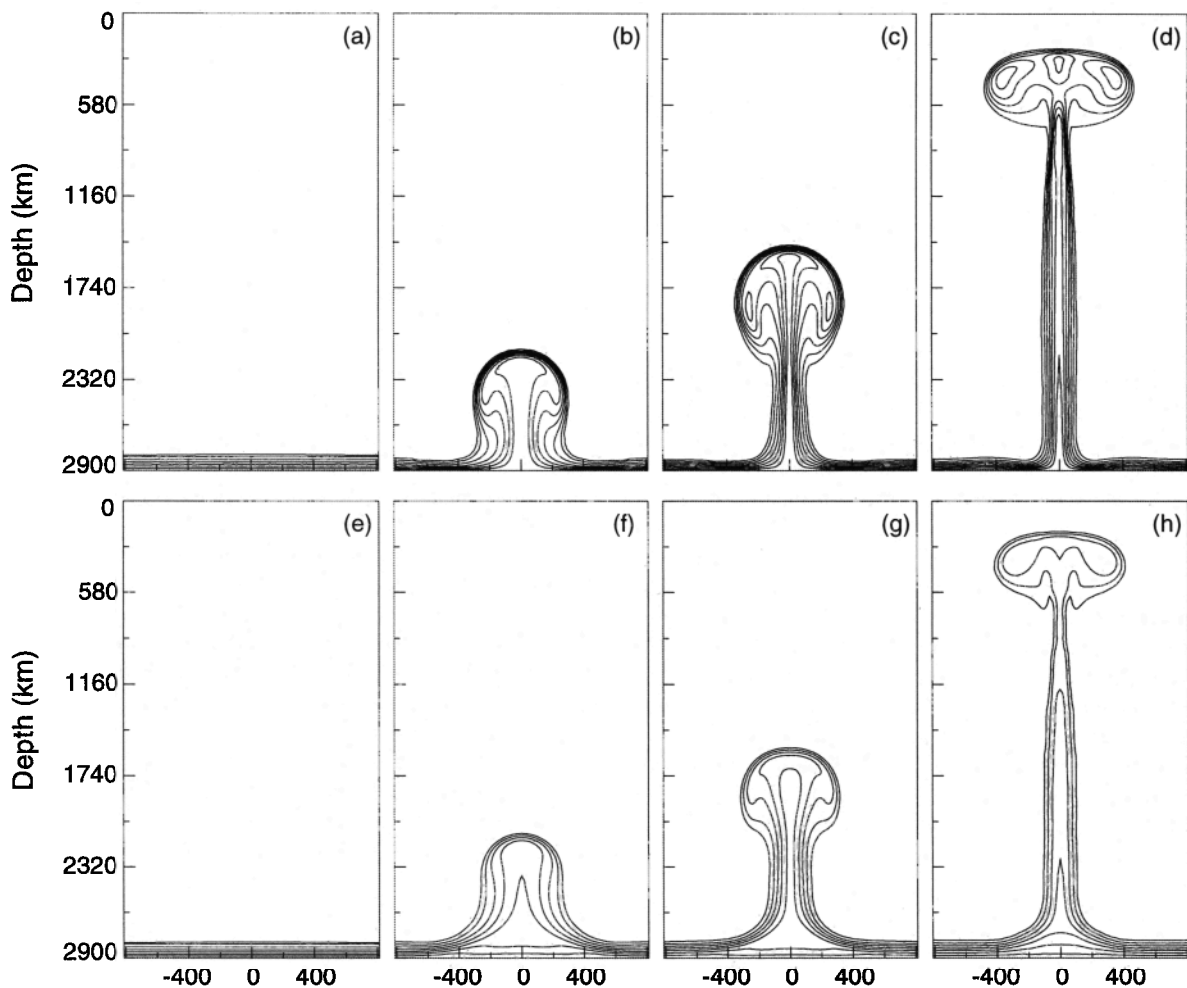


Figure 1. Evolution of a mantle plume for the reference model ($\Delta T_{bl}=800^\circ\text{C}$). Contours of excess potential temperature every 100°C . Top: $\Delta\rho_{ch}=0\%$. Bottom: $\Delta\rho_{ch}=5\%$. Horizontal axis: radial distance (km), the total radius (not shown) is 1500 km.

if $\alpha=1 \times 10^{-5} \text{ } ^\circ\text{C}^{-1}$ at the base of the mantle). The initial thickness of the thermal boundary layer is calculated as $\delta=K\Delta T_{bl}/Q$, $K=4.2 \text{ W m}^{-1} \text{ } ^\circ\text{K}^{-1}$ is the thermal conductivity and Q is the heat flow rate by convection calculated from the relation between the Nusselt number ($Nu=Q/Q_c$) and the Rayleigh number $Nu=0.225Ra^{-1/3}$ [see *Farnetani and Richards*, 1995]. For the reference model the calculated core heat flux is $5 \times 10^{12} \text{ W}$ (i.e., less than 20% of the surface heat flux). Throughout the chemical layer, 30 km thick, the C-field has initial value $C=1$, (i.e., 100% of chemically distinct material), elsewhere $C=0$. The C-field is used to calculate the excess chemical density of each fluid parcel during the flow $\Delta\rho_c(r,z,t)=\rho_0\beta C(r,z,t)$ (e.g., given $\beta=0.05$, $C=1$ and $C=0.2$ correspond, respectively, to a material 5% and 1% chemically denser than the surrounding mantle). No influx of dense material is allowed through the boundaries. This is probably a reasonable approximation since for all models the time scale of plume formation and rise is less than 140 My. At the surface the vertical velocity component is $U_z=0$, while the radial velocity is set to $U_r=0$ from the surface to 40 km depth in order to model a non-rifting lithosphere. On the bottom boundary $U_z=0$ and $\delta U_r/\delta z=0$ (radial free slip), while along the axis of symmetry and the right side of the box $U_r=0$ and $\delta U_z/\delta r=0$ (vertical free slip). Each rectangular element of the non-uniform Eulerian grid has radius $13 < r_e < 19 \text{ km}$ and height $13 < z_e < 19 \text{ km}$. Smaller elements are closer to the axis of symmetry and at the boundaries.

Results

The model parameters vary about a reference case with $\Delta T_{bl}=800^\circ\text{C}$. Figure 1 (top) shows the time evolution of a plume for the case without chemical buoyancy ($\beta=0$). For the initial condition (fig. 1a) the excess potential temperature across the thermal boundary layer decreases as an error function conduction profile. After 36 My the plume has developed a large head (ca. 300 km radius) and a narrow tail (fig. 1b). After 41 My (fig. 1c) the highest excess temperatures (600-700°C) are along the axis and within the plume head in a torus shaped region. After 56 My (fig. 1d) the plume head is in the upper mantle and starts to spread beneath the lithosphere (ca. 120 km thick, not shown in the figure). The plume head does not cool significantly during its ascent from the source region to the upper mantle, and the calculated excess temperatures (450-500°C) are far too elevated compared to estimates of excess temperature for flood volcanism. Figure 1 (bottom) shows the time evolution of a plume for the case with a 30 km thick chemical layer, 5% denser than the overlying mantle ($\beta=0.05$). Since the chemical field has a modest numerical diffusion, $\Delta\rho_{ch}$ is set to zero also for $C < 0.1$. This is a conservative assumption that avoids assigning chemical buoyancy to fluid parcels those low C-field may be the effect of numerical diffusion. The integrated volume of the chemically denser material remains approximately constant during plume ascent ($V_{final}=97\% V_{initial}$). The initial condition for the temperature field (fig. 1e) is identical to the previous case. The ascent of the hotter fluid at the base of the boundary layer is hindered by the presence of the chemical layer (fig 1f, after 65 My). A secondary circulation develops along the axis and material 100-150 km above the bottom boundary is entrained downward by the counterrotating flow, similarly to what observed by *Kellogg and King* [1993]. The plume head in

the lower mantle has an excess temperature of ca. 400°C (fig 1g, after 75 Ma). Note that elapsed time is given only for comparison and that the time scales would be shorter for stronger temperature dependence of viscosity. The final excess temperature of the plume head spreading beneath the lithosphere (fig 1h) is 300-350°C.

Figure 2a shows the average plume head excess temperature (ΔT_h) as a function of chemical density for the reference model with $\Delta T_{bl}=800^\circ\text{C}$ (i.e., $1020 < \Delta T_{rbl} < 1300^\circ\text{C}$), for the model with $\Delta T_{bl}=500^\circ\text{C}$ ($630 < \Delta T_{rbl} < 810^\circ\text{C}$) and for the model with $\Delta T_{bl}=1100^\circ\text{C}$ ($1400 < \Delta T_{rbl} < 1780^\circ\text{C}$). For all models ΔT_h is calculated by averaging the plume head excess temperature over an area of 650 km diameter at a constant depth (350 km). The results indicate that the presence of a chemically denser layer reduces considerably the excess temperature of plume heads. Figure 2b shows the same results, but now the plume head excess temperature is normalized over the initial excess temperature across the thermal boundary layer, and the excess chemical density ($\Delta\rho_{ch}$) is normalized over the thermal density deficiency ($\Delta\rho_t$). For all models with $\Delta\rho_{ch}=0$ the calculated ΔT_h is $0.6\Delta T_{bl}$, while for $\Delta\rho_{ch}=3\Delta\rho_t$ the calculated ΔT_h is reduced to approximately $0.3\Delta T_{bl}$.

A further set of models investigates the effect of a 60 km thick chemical boundary layer. In such case plume ascent is considerably hindered and the final plume head excess temperature ΔT_h is less than $0.25\Delta T_{bl}$ for $\Delta T_{bl}=800^\circ\text{C}$.

Figure 3 shows the excess potential temperature across the plume tail at 2000 km depth, for models with $\Delta T_{bl}=800^\circ\text{C}$. The excess temperature is up to 700°C for $\Delta\rho_{ch}=0\%$, while for $\Delta\rho_{ch}=5\%$ the temperatures are lower (ca. 400°C). For comparison, plume tail excess temperature of 300°K were estimated in the lower mantle beneath the weak Bowie hotspot [*Nataf and vanDecar* 1993]. Both the excess temperature and the buoyancy flux across the tail decrease by increasing the chemical density contrast across the source region.

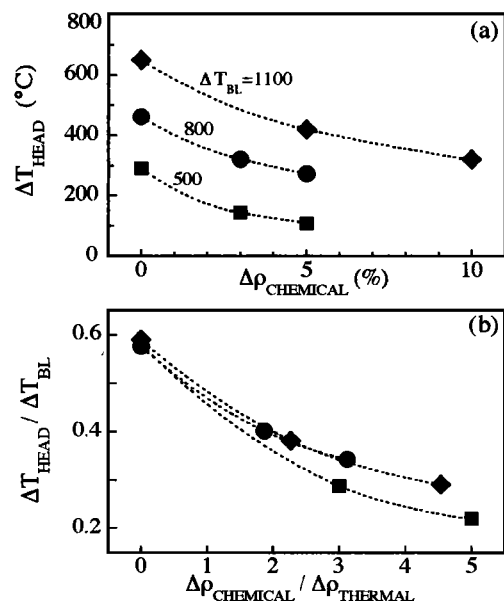


Figure 2. (a) Excess chemical density vs. plume head excess temperature (ΔT_{head}) for models with initial boundary layer excess temperature $\Delta T_{bl}=500^\circ\text{C}$ (squares), $\Delta T_{bl}=800^\circ\text{C}$ (circles), $\Delta T_{bl}=1100^\circ\text{C}$ (diamonds). (b) Excess chemical density normalized over thermal density vs. $\Delta T_{head}/\Delta T_{bl}$.

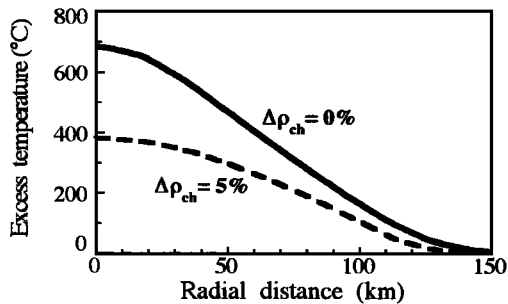


Figure 3 Radial distance vs. excess temperature across the plume tail at 2000 km depth for models with $\Delta T_{bl}=800^\circ\text{C}$.

Discussion

The existence of chemical heterogeneities at the base of the Earth's mantle is still a matter of debate, however, several arguments favour a density increase across D". Seismological observations of a sharp decrease in seismic velocities above the core mantle boundary may support the existence of a density increase, probably due to an iron enrichment. Dynamical models show that D" must be a chemical layer to prevent its entrainment by overlying convecting mantle [e.g., *Davies and Gurnis*, 1986; *Sleep*, 1988]. The origin of a chemical layer may be attributed to: 1) Differentiation processes [see *Manga and Jeanloz*, 1996 and references therein]. 2) Reaction processes between mantle silicates and iron from the outer core. Mineral physics experiments at high pressure show that molten iron preferentially dissolves FeO and penetrates into the silicate [*Knittle and Jeanloz*, 1989]. Although the infiltration of core alloy into the base of the lower mantle is possible only over distances of order 1-100 m [*Poirier and Le Mouél*, 1992], convection may sweep the infiltrated layer upward [*Knittle and Jeanloz*, 1991]. 3) Partial melting would generate high density liquids if iron partitions preferentially into the melt relative to the coexisting solids. *Williams and Garnero* [1996] have attributed a 10% decrease in compressional wave velocity over a 40 km thick zone at the base of the mantle to the presence of partial melts.

In this paper I investigated the role of a thin chemical layer across D" on the dynamics of plume formation and ascent. I have found that if the source region is *only* a thermal boundary layer, the calculated plume head excess temperatures in the uppermost mantle are too high. For models with a 30 km thick chemical layer 5% denser than the surrounding mantle, the average excess temperature across the plume head is plausible. The role of the chemical layer is to effectively prevent the lower part of the thermal boundary layer from becoming part of the plume. These results indicate that a chemically denser layer at the base of the Earth's mantle may have a fundamental role in governing plume temperature anomalies. However, a trade-off between the excess density and the thickness of the chemical layer hampers a definitive conclusion which may be reached only through a profound understanding of the nature and origin of chemical heterogeneities across D".

Acknowledgments. I thank E. Garnero, P. vanKeken and an anonymous reviewer for thoughtful reviews, M. Richards, C. Lithgow-Bertelloni and L. Stixrude for useful comments that improved the manuscript. I also thank H-C. Nataf, P. Cardin and L. Fleitout for encouragement and support. The computer resources were provided by

IBM, at Institut Pierre-Simon Laplace. This research was supported by a CNRS fellowship.

References

- Boehler, R., A. Chopelas, and A. Zerr, Temperature and chemistry of the core-mantle boundary, *Chemical Geology*, 120, 199-205, 1995.
- Davies, G. F., and M. A. Richards, Mantle convection, *J. of Geology*, 100, 151-206, 1992.
- Davies, G. F., and M. Gurnis, Interaction of mantle drags with convection: lateral heterogeneity at the core-mantle boundary, *Geophys. Res. Lett.*, 13, 1517-1520, 1986.
- Farnetani, C. G., and M. A. Richards, Numerical investigation of the mantle plume initiation model for flood basalt events, *J. Geophys. Res.*, 99, 13813-13833, 1994.
- Farnetani, C. G., and M. A. Richards, Thermal entrainment and melting in mantle plumes, *Earth Planet. Sci. Lett.*, 136, 251-267, 1995.
- Garnero, E. J., and D. V. Helmberger, Seismic detection of a thin laterally varying boundary layer at the base of the mantle beneath the central Pacific, *Geophys. Res. Lett.*, 23, 977-980, 1996.
- Griffiths, R. W., and I. H. Campbell, Stirring and structure in mantle starting plumes, *Earth Planet. Sci. Lett.*, 99, 66-78, 1990.
- Jeanloz, R., and S. Morris, Temperature distribution in the crust and in the mantle, *Ann. Rev. Earth Planet. Sci.*, 14, 377-415, 1986.
- Kellogg, L. H., and S. D. King, Effect of mantle plumes on the growth of D" by reaction between the core and the mantle, *Geophys. Res. Lett.*, 20, 379-382, 1993.
- Knittle, E., and R. Jeanloz, Simulating the core-mantle boundary: an experimental study of high pressure reactions between silicates and liquid iron, *Geophys. Res. Lett.*, 16, 609-612, 1989.
- Knittle, E., and R. Jeanloz, Earth's core-mantle boundary: Results of experiments at high pressures and high temperatures. *Science*, 251, 1438-1443, 1991.
- Loper, D., and T. Lay, The core-mantle boundary region, *J. Geophys. Res.*, 100, 6379-6420, 1995.
- Manga, M., and R. Jeanloz, Implications of a metal-bearing chemical boundary layer in D" for mantle dynamics, *Geophys. Res. Lett.*, 23, 3091-3094, 1996.
- Nataf, H-C., and J. vanDecar, Seismological detection of a mantle plume? *Nature*, 364, 115-120, 1993.
- Poirier, J-P., and J-L. Le Mouél, Does infiltration of the core material into the lower mantle affect the observed geomagnetic field? *Phys. Earth Planet. Inter.*, 73, 29-37, 1992.
- Richter, F. M., and D. P. McKenzie, On some consequences and possible causes of layered mantle convection, *J. Geophys. Res.*, 86, 6124-6133, 1981.
- Schilling, J-G., Fluxes and excess temperatures of mantle plumes inferred from their interaction with migrating mid-ocean ridges, *Nature*, 352, 397-403, 1991.
- Sleep, N. H., Gradual entrainment of a chemical layer at the base of the mantle by overlying convection, *Geophys. J.*, 95, 437-447, 1988.
- Stacy, F. D., and D. E. Loper, The thermal boundary layer interpretation of D" and its role as a plume source, *Phys. Earth Planet. Inter.*, 33, 45-55, 1983.
- vanKeken, P.E., S. D. King, H. Schmeling, U. R. Christensen, D. Neumeister, and M.-P. Doin, A comparison of methods for the modeling of thermochemical convection, in press *J. Geophys. Res.*, 1997.
- Vinnik, L., B. Romanowicz, Y. Le Stunff, and L. Makeyeva, Seismic anisotropy in the D" layer, *Geophys. Res. Lett.*, 22, 1657-1660, 1995.
- Boehler, R., A. Chopelas, and A. Zerr, Temperature and chemistry of the core-mantle boundary, *Chemical Geology*, 120, 199-205, 1995.
- Watson, S., and D. P. McKenzie, Melt generation by plumes: A study of Hawaiian volcanism, *J. Petrol.*, 12, 501-537, 1991.
- Williams, Q., and E. J. Garnero, Seismic evidence for partial melt at the base of the Earth's mantle, *Science*, 273, 1528-1530, 89-109, 1996.

Cinzia G. Farnetani, Lab. de Dynamique des Systèmes Géologiques, Institut de Physique du Globe, 4, Place Jussieu, 75252 Paris France.

(Received: January 14, 1997; Accepted: April 1, 1997)