Generation of mega-plumes from the core-mantle boundary in a compressible mantle with temperature-dependent viscosity

Paul F. Thompson [1] and Paul J. Tackley

Department of Earth and Space Sciences, University of California, Los Angeles

Abstract. We have investigated the development of axisymmetric mantle plumes in a compressible mantle with temperature-dependent viscosity, using two different values for the thermal activation energy. Thermodynamic parameters were based on realistic Earth values in the shallow mantle. We assumed a compressible mantle with a phase transition and a two layer viscosity structure (mantle 30 times more viscous beneath the phase change). A dramatic transition in plume dynamics occurred as the thermal activation energy was increased from 250 to 500 kJ mol^{-1}. A "mega-plume" was formed from the merging of small-scale instabilities swept into the plume's axis, rapidly flushing the entire contents of the hot thermal boundary layer into the upper mantle. Hot plume material with a temperature anomaly of 250 K reached the lithosphere for our assumed temperature drop of 500 K across the base of the mantle. Viscous heating over four orders of magnitude greater than chondritic radiogenic sources was observed during the initial plume-lithosphere interaction.

Introduction

Hot plumes arising from a thermal boundary layer deep in the mantle are widely believed to be an important feature of the Earth's mantle dynamics, and may be the key to understanding lithospheric structures such as hot spots and flood basalts [*Morgan*, 1971]. Flood basalts are seen as an expression of the strong initial plume head, while longer lived hot spots are associated with the subsequent plume conduit [e.g. *Richards et al.*, 1989; *White and McKenzie*, 1995]. With a temperature drop potentially as high as 1500 K [e.g. *Boehler*, 1996; *Christensen*, 1984], the thermal boundary layer above the core-mantle interface may be the deep mantle source for plumes.

Our current understanding of mantle plume dynamics is based on simplified models which generally make one of two assumptions: constant viscosity or Boussinesq (incompressible) conditions. While there have been detailed plume studies that include compressibility [*Zhao and Yuen*, 1987] and other studies that include temperature-dependent viscosity [*Davies*, 1995; *Kellogg and King*, 1997; *Schubert et al.*, 1995], both complexities have only been included in steady-state or weakly time-dependent models [*Albers and Christensen*, 1996]. The effects of compressibility are two-fold, introducing depth-dependent material properties; and introducing the additional terms of viscous dissipation and adiabatic heating into the energy equation. Three-dimensional, whole-mantle convection models have given some insights: the depth-dependent properties associated with compressibility produce slower, broader plumes [*Balachandar et al.*, 1992; *Tackley*, 1996], while temperature-dependent viscosity acts the opposite way, invigorating the flow and encouraging smaller scales with higher velocities [*Balachandar et al.*, 1995; *Tackley*, 1994].

Here the initiation and propagation of mantle plumes from the base of the mantle is studied. A high-resolution local model is used, allowing for much more realistic material properties, such as viscosity variation, than have been possible either in global 3-D models or local plume studies. We focus particularly on the effects which result from the combination of compressibility and strongly temperature-dependent viscosity. This is achieved with two simulations that use different thermal activation energies for the viscosity law. In the lower activation energy case, $E_{act} = 250$ kJ mol^{-1}, the viscosity contrast across the lower thermal boundary layer is 150, while the higher activation energy, $E_{act} = 500$ kJ mol^{-1}, produces viscosity contrasts of up to 22,500.

Model and Method

Calculations were performed using STAG3D, which is a three-dimensional, Cartesian code using a finite-volume, multigrid technique, here adapted to model two-dimensional, axisymmetric, cylindrical geometry. Details of the compressible anelastic-liquid equations, reference state, and numerical implementation are discussed fully in *Tackley* [1996]; we discuss here only the salient features of the model as they pertain to this study.

An axisymmetric domain is assumed, with a radius equal to the depth of the domain ($D = 3000$ km). For computational convenience, the top cold thermal boundary layer is removed, and a rigid boundary condition represents the lithosphere. The bottom boundary is free-slip and isothermal, and the side boundaries are reflecting. Temperature is fixed at the top as $T_0 = 1500K$, with a superadiabatic temperature difference of $\Delta T_{sa} = 500K$ across the mantle.

The derivation of this reference state is as described in *Tackley* [1996], except for the addition here of a discontinu-

[1] Now at Center for Space Research, The University of Texas at Austin.

Paper number 98GL01228.
0094-8534/98/98GL-01228$05.00

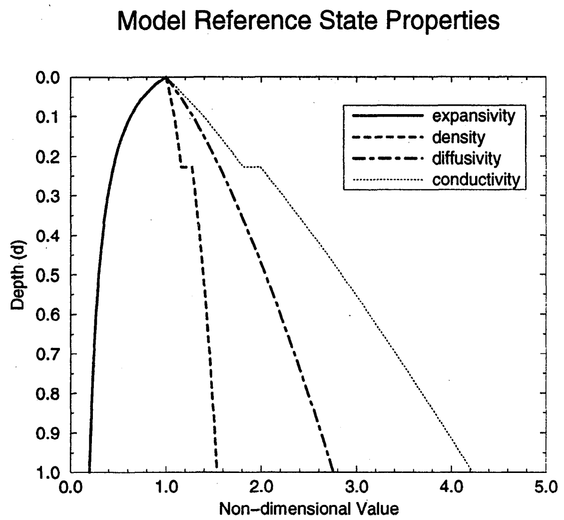

Figure 1. Variation of the non-dimensional reference state thermodynamic parameters of thermal expansivity, density, thermal diffusivity, and thermal conductivity.

ous increase in density, $\delta\rho = 400$ kg m^{-3}, from the upper to lower mantle due to an endothermic phase change at 660 km depth. Shown in Figure 1 is the variation of the thermodynamic parameters with depth. These parameters are non-dimensionalized by representative values in the upper mantle. The resultant Rayleigh number, $Ra_0 = 2.43 \times 10^7$, is based on these top boundary values and is a primary input parameter for the model.

A modified Arrhenius law is used to describe the variation of viscosity with temperature,

$$\eta(T_{sa}) = A_0 exp\left(\frac{E_{act}}{R(T_{sa}+T_0)}\right), \quad (1)$$

where R is the gas constant, T_{sa} is superadiabatic temperature, T_0 is top boundary temperature, and E_{act} is thermal activation energy. A jump in the background viscosity by a factor of 30 is assumed at 660 km depth; therefore, A_0 is defined such that $\eta(T_{sa} = 0) = \eta_0$ in the upper mantle and $30\eta_0$ in the lower mantle.

First, we consider an activation energy of 250 kJ mol^{-1}. The second simulation is identical to the first, except that $E_{act} = 500$ kJ mol^{-1}. In the upper mantle, where deformation may be dominated by dislocation creep, a suitable value for the thermal activation energy may be as large as 540 kJ mol^{-1}. However, in the lower mantle, where diffusion creep may be the dominant deformation mechanism, a reasonable value may be in the range of 240-300 kJ mol^{-1} [e.g., see *Karato and Wu*, 1993]. Due to these uncertainties, combined with computational limitations, mantle convection models commonly assume a simplified rheology with a reduced effective activation energy [e.g. *Davies*, 1995].

The initial mantle temperature state is adiabatic, and includes a temperature step across the 660 km discontinuity due to latent heat associated with the endothermic phase change. A complementary error function is used to define the temperature structure of an initial bottom thermal boundary layer 18 km thick. It completely contains the superadiabatic temperature drop, ΔT_{sa}. This is the smallest resolvable boundary layer for our choice of a numerical grid, and it requires that the boundary layer conductively thicken before convective instability can occur naturally. This thin layer could arise in the mantle under certain conditions, such as in the case of a subducted slab deposited upon the core-mantle boundary (CMB), sweeping aside the less viscous, resident hot material. Our experimental conditions allow for a more extensive plume head than is allowed in other plume models which include a heated patch at the core-mantle boundary [*Kellogg and King*, 1997; *Schubert et al.*, 1995].

Resolutions along the horizontal and vertical axes vary from 128 x 128 grid points in the reduced activation energy case, to 512 x 256 in the higher activation energy case. The grid is non-uniform with a vertical grid refinement by about a factor of three in the upper and lower boundaries, and at 660 km depth.

Results

Reduced Activation Energy

The development of the plume using an activation energy of $E_{act} = 250$ kJ mol^{-1} is shown in Figure 2. Times are measured relative to when the axial instability begins to rise above the boundary layer (see first frame of Figs. 2 and 3). The initial conductive thickening of the boundary layer was followed by the separation of an instability at the axis of symmetry. A distinct head and tail were observed, similar to a cavity plume. The head of the plume was pinched and partially deflected by the 660 km phase change, but easily penetrated into the upper mantle. It continued to rise, reaching the lithosphere approximately 96 Myr after initiation, similar to the rise times found in previous Boussinesq calculations [e.g. *Christensen*, 1984].

The plume head was 500 km in radius, and retained a 250 K temperature anomaly relative to the upper mantle upon initially reaching the lithosphere. Subsequently, the plume spread out under much of the lithosphere, reaching the edge of the domain at 170 Myr. Largely as a consequence of adiabatic cooling, the excess temperature of a plume is commonly observed to be substantially less than the temperature drop prescribed across the lower boundary layer [*Albers and Christensen*, 1996; *Zhao and Yuen*, 1987]. However, a 250 K temperature anomaly is at the high end of the 150-250 K temperature range necessary for generating the types of melt present during the formation of flood basalts [*White and McKenzie*, 1995].

Large values of the viscous dissipation term were observed in the regions where the rising plume interacted with the base of the lithosphere and endothermic phase change [*Balachandar et al.*, 1995]. The maximum value of viscous heating near the lithosphere was 10 K Myr^{-1}, while a peak value of 30 K Myr^{-1} occurred as the plume head first interacted with the phase change. A secondary maximum was observed throughout the simulation alongside the plume conduit. For comparison, the radiogenic heating rate in carbonaceous chondrites given by *Stacey* [1992] as $H = 5.2 \times 10^{-12}$W kg^{-1}, corresponds to 0.135 K Myr^{-1} with our model parameters.

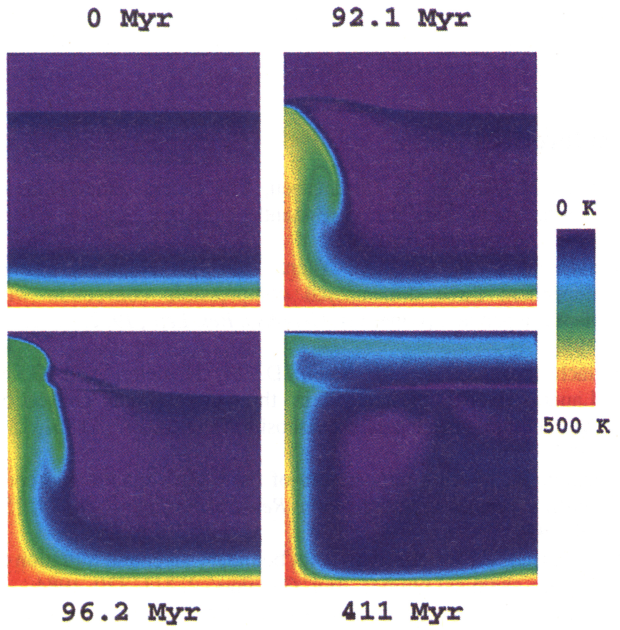

Figure 2. Development of plume case with $E_{act} = 250$ kJ mol^{-1}. Superadiabatic temperature.

The simulation was continued until 411 Myr after initiation, at which point non-axisymmetric instabilities begin to form in the D″-layer. The longer-lived plume conduit that remained was reduced in size to approximately 250 km radius in the lower mantle. However, the assumption of a fixed, isothermal top boundary for such long time periods is not realistic.

Preferred Activation Energy

Shown in Figure 3 is a plume case using a value of $E_{act} =$ 500 kJ mol^{-1}, producing viscosity contrasts up to 22,500 across the initial plume. The dynamics of this plume are dramatically different from the lower activation energy case. As the thermal boundary layer conductively thickens, the local *Ra* is exceeded because of the low viscosities within, producing small-scale convection in the boundary layer before the large instability grows out of the boundary layer [previously noted by *Christensen*, 1984]. The small-scale instabilities moved towards the main axial plume, supplying it with a large volume of hot material, forming a very broad (up to 1000 km radius), well-mixed, low-viscosity plug. This "mega-plume" rapidly flushed the entire hot lower boundary layer into the upper mantle, a potential mechanism for the creation of a superplume event as discussed by *Larson* [1991]. Other calculations (not shown here) with a thicker initial boundary layer displayed an even more dramatic mega-plume effect due to the greater amount of source material for the initial plume pulse.

The behavior in this case is thus caused by two factors: strong temperature-dependent viscosity which allows localized convection and strong advection within the boundary layer, and depth-dependent viscosity and other properties which slow the development of the large-scale plume. Similar Boussinesq calculations we performed were not able to reproduce this effect. Furthermore, this type of mega-plume behavior has not been seen in planar 2-D compressible Cartesian calculations, probably because an axisymmetric plume is able to tap a larger volume of source material relative to its size as material converges radially towards the plume conduit.

As the simulation progressed beyond 96 Myr, material rising up the plume conduit had greater difficulty penetrating the phase change at this higher activation energy. Plume material pooled just below the phase change boundary at 660 km depth, forming a hot layer which produced small, intermittent plumes (actually "rings" due to the axisymmetry). This partial and intermittent penetration could provide a mechanism for a hot spot to have variation in the plume flux, while being fed from a relatively stable thermal bound-

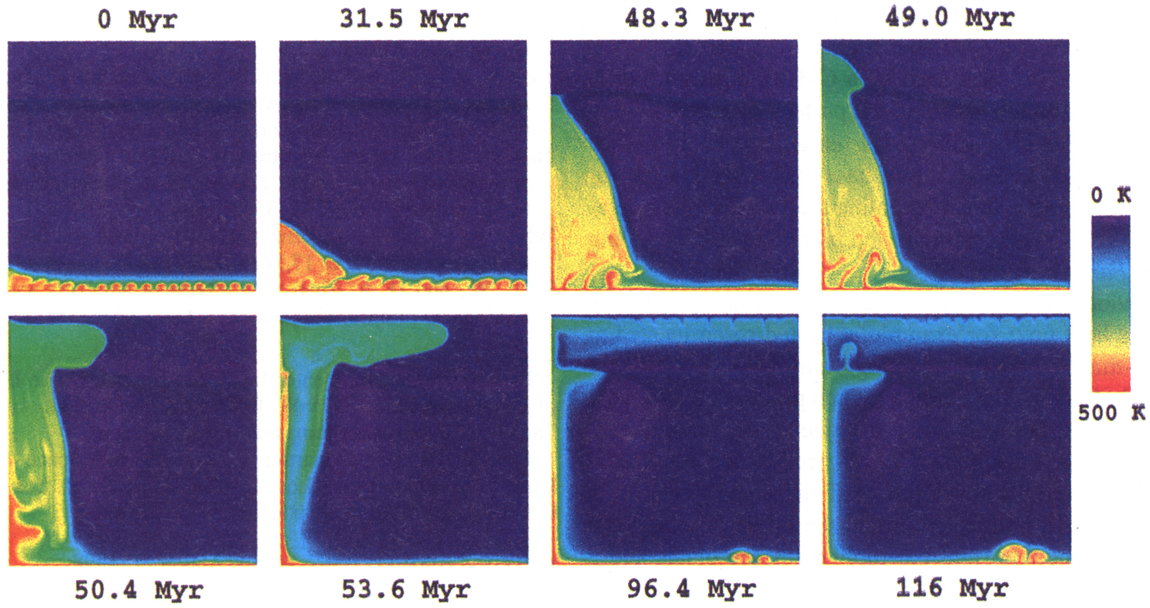

Figure 3. Development of plume case with $E_{act} = 500$ kJ mol^{-1}. Superadiabatic temperature.

ary layer at the CMB. Full 3-D geometry will be required to properly evaluate this effect.

Viscous dissipation was highly localized in space and time, with very large peak values over three orders of magnitude greater than chondritic heating rates (500 K Myr^{-1} occurring during the initial plume-lithosphere interaction). This is significantly larger than the magnitudes of viscous dissipation obtained by *Balachandar et al.* [1995] (values on the order of 20-30 times greater than chondritic sources). With the higher viscosity contrasts and Rayleigh numbers used in our study, convection is more vigorous and greater magnitudes of viscous dissipation are to be expected.

Conclusions and Discussion

A transition from the conventional plume behavior (plume head and tail structure) to a broad scale mega-plume (well-mixed, low viscosity plug) took place for geophysically reasonable physical parameters. Future work should consider other activation energies in order to more clearly define this transition. Large peak values of viscous dissipation occurred in the upper mantle, possibly aided by the positive feedback between viscous dissipation and temperature-dependent viscosity which is known to produce faster, more vigorous plumes [*Zhao and Yuen*, 1987]. The mega-plume produced in this study rises significantly faster than a conventional plume, contrary to the expectation that the background viscosity controls the rise velocity as given by Stokes flow [*Christensen*, 1984]. In slightly more than 50 Myr the entire volume of the CMB boundary layer was flushed into the shallow mantle, an event which is likely to have dramatic geophysical consequences [e.g. *Larson*, 1991]. The remnant plume conduit is very narrow and may be difficult to detect in the geoid or with seismic tomography. However, this model predicts the formation of a warm, low viscosity layer below the transition zone which may be detectable [*Kido and Cadek*, 1997].

Despite the transition in plume behavior, the temperature anomaly in the upper mantle was relatively constant, consistent with *Albers and Christensen* [1996]. A 250 K temperature anomaly, was produced by a CMB temperature drop of only 500 K, much less than the preferred range of 1000-1500 K [e.g. *Boehler*, 1996; *Christensen*, 1984]. It may be that the D″-layer observed at the base of the mantle is chemically distinct and stable, preventing the full CMB temperature drop from participating in the production of upwelling plumes [*Farnetani*, 1997].

In previous plume studies it has been common to enforce a specific heat flux; for example, a heated patch located at the core-mantle boundary [*Kellogg and King*, 1997; *Schubert et al.*, 1995]. While this approach produces a relatively stable plume conduit, it limits the size of the plume head. Clearly, a range of boundary layers needs to be considered in order to model the development of an initiating plume. In order to assess a plume's effect on the lithosphere, models will need to include rigid but mobile surface plates and a low viscosity asthenosphere.

Acknowledgments. We would like to thank S. D. King and A. M. Leitch for their constructive reviews. This research was supported by NSF grant EAR9506683.

References

Albers, M., and U. R. Christensen, The excess temperature of plumes rising from the core-mantle boundary., *Geophys. Res. Lett.*, *23*, 3567–3570, 1996.

Balachandar, S., D. A. Yuen, and D. Reuteler, Time-dependent three dimensional compressible convection with depth-dependent properties, *Geophys. Res. Lett.*, *19*, 2247–2250, 1992.

Balachandar, S., D. A. Yuen, D. M. Reuteler, and G. S. Lauer, Viscous dissipation in three-dimensional convection with temperature-dependent viscosity, *Science*, *267*, 1150–1153, 1995.

Boehler, R., Melting temperature of the Earth's mantle and core: Earth's thermal structure, *Annu. Rev. Earth Planet. Sci.*, *24*, 15–40, 1996.

Christensen, U., Instability of a hot boundary layer and initiation of thermo-chemical plumes, *Ann. Geophys.*, *2*, 311–320, 1984.

Davies, G. F., Penetration of plates and plumes through the mantle transition zone, *Earth Planet. Sci. Lett.*, *133*, 507–516, 1995.

Farnetani, C. G., Excess temperature of mantle plumes: the role of chemical stratification across D″, *Geophys. Res. Lett.*, *24*, 1583–1586, 1997.

Karato, S., and P. Wu, Rheology of the upper mantle: A synthesis, *Science*, *260*, 771–778, 1993.

Kellogg, L. H., and S. D. King, The effect of temperature dependent viscosity on the structure of new plumes in the mantle: Results of a finite element model in a spherical, axisymmetric shell, *Earth Planet. Sci. Lett.*, *148*, 13–26, 1997.

Kido, M., and O. Cadek, Inferences of viscosity from the oceanic geoid: Indication of a low viscosity zone below the 660-km discontinuity, *Earth Planet. Sci. Lett.*, *151*, 125–137, 1997.

Larson, R. L., Latest pulse of Earth: Evidence for a mid-Cretaceous superpume, *Geology*, *19*, 547–550, 1991.

Morgan, J. W., Convection plumes in the lower mantle, *Nature*, *230*, 42–43, 1971.

Richards, M. A., R. A. Duncan, and V. E. Courtillot, Flood basalts and hot-spot tracks: Plume heads and tails, *Science*, *246*, 103–107, 1989.

Schubert, G., C. Anderson, and P. Goldman, Mantle plume interaction with an endothermic phase change, *J. Geophys. Res.*, *100*, 8245–8256, 1995.

Stacey, F. D., *Physics of the Earth*, 3rd ed., Brookfield, Kenmore, Queensland, Australia, 1992.

Tackley, P. J., Three-dimensional models of mantle convection: Influence of phase transitions and temperature-dependent viscosity, Ph.D. thesis, California Institute of Technology, 1994.

Tackley, P. J., Effects of strongly variable viscosity on three-dimensional compressible convection in planetary mantles, *J. Geophys. Res.*, *101*, 3311–3332, 1996.

White, R. S., and D. McKenzie, Mantle plumes and flood basalts, *J. Geophys. Res.*, *100*, 17,543–17,585, 1995.

Zhao, W., and D. A. Yuen, The effects of adiabatic and viscous heatings on plumes, *Geophys. Res. Lett.*, *14*, 1223–1226, 1987.

Paul F. Thompson, Center for Space Research, The University of Texas at Austin, 3925 W. Braker Ln., Suite 200, Austin, TX 78759-5321. (e-mail: thompson@csr.utexas.edu)

Paul J. Tackley, Department of Earth and Space Sciences, University of California, Los Angeles, Box 951567, Los Angeles, CA 90095-1567. (e-mail: ptackley@ess.ucla.edu)

(Received April 11, 1997; revised March 11, 1998;
accepted March 20, 1998.)