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The Role of Rheology in Lithospheric Thinning by Mantle Plumes

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Abstract. Lithospheric thinning by mantle plumes is an important planetary heat transfer process resulting in the broad topographic uplift that characterizes the volcanic rises on Venus, the Tharsis rise on Mars, and several hotspot swells on Earth. We present a suite of time-dependent, three-dimensional numerical calculations of a plume impinging upon the lithosphere in a temperature-dependent viscosity mantle. Efficient lithospheric thinning is found to depend on the formation of convective instabilities in the plume-lithosphere boundary layer. These instabilities are non-axisymmetric, time-dependent, and have horizontal scales of a few tens of kilometers. These instabilities depend on the temperature-dependence of viscosity and occur when the plume's viscosity is about an order of magnitude less than the background mantle, as predicted by boundary-layer theory. Thus, in planetary mantles, plumes with excess temperatures of 100 to 200 K will efficiently thin the lithosphere via small-scale convective instabilities.

Introduction

Hotspots are associated with broad regions (1000 km or more) of uplift in a wide array of settings including fast oceanic plates such as the Pacific, slow plates such as the African, and single-plate lithospheres such as Venus and Mars. Mantle plumes are widely believed to be responsible for both the increased volcanism and the uplift observed at hotspots [*Wilson*, 1963]. On the fast moving Pacific plate, the similarity of the subsidence of the Hawaiian swell topography to the subsidence of the oceanic lithosphere with age prompted *Crough* [1978] to propose that the lithosphere was "rejuvenated" or thinned as it passed over the plume.

Lithospheric thinning by mantle plumes has been much debated since then, with both the observational basis for thinning and the plausibility of large thinning rates (required to explain the rapid rise of the Hawaiian swell) coming into question. Numerical models in a variety of geometries [*Robinson et al.*, 1987; *Monnereau et al.*, 1993; *Ribe and Christensen*, 1994] suggested that swells were supported by dynamic pressure variations rather than lithospheric thinning, although *Parsons and Daly* [1983] had shown that convective models for swell formation only differ in the locations of temperature variations and the dynamic stresses required to maintain them (lithosphere vs. asthenosphere). It was also proposed that the Hawaiian swell may be the result of thermal and chemical buoyancy with [*Moore and Schubert*, 1997a] or without [*Phipps-Morgan et al.*, 1995] thinning. An analytical study of upwellings revealed that thinning is slow above broad (500 km) upwellings, but rapid over a 35 km upwelling [*Davies*, 1994]. Models in which the plume was represented by a high temperature or high heat flow boundary condition on the lithosphere [*Spohn and Schubert*, 1983] and semi-analytical models of plume flow in the asthenosphere [*Sleep*, 1994] demonstrated that conduction was too slow and that convection needed to supply several times the normal mantle heat flow for rapid thinning. A study using mean-field theory [*Yuen and Fleitout*, 1985] showed that small-scale convective instabilities were capable of providing the necessary excess heat flow.

Laboratory studies [*Griffiths and Campbell*, 1991; *Davaille and Jaupart*, 1994], demonstrated that such instabilities obey scaling laws derived from boundary layer theory of convection in temperature-dependent viscosity fluids [*Morris and Canright*, 1984; *Solomatov*, 1995]. Recently, we demonstrated that these scalings also apply to instabilities generated by a plume interacting with a moving plate in a three-dimensional numerical model [*Moore et al.*, 1998].

In this report we present the results of several numerical experiments in which a plume interacts with a stationary plate. By varying the temperature of the plume and the temperature dependence of the viscosity, we determine the criterion for small-scale instability in the plume-lithosphere boundary layer. Since the plate is motionless with respect to the plume source, we are able to compare with axisymmetric results.

Model

We performed our numerical experiments in a three-dimensional box 400 km deep by 3200 km on a side at the top of the mantle. Within this box the equations of mass, momentum, and energy conservation in an infinite Prandtl number, Boussinesq fluid are solved numerically in three dimensions by finite volume discretization on a staggered grid using STAG3D [*Tackley*, 1996]. The surface is rigid and isothermal, the side

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boundaries are reflecting, and the bottom boundary is permeable, where a plume conduit is modeled by a hot patch with a Gaussian temperature profile. Hot material entering in the region of the conduit forms a plume. In this study, the plume conduit has a constant half-width of 67 km and a peak excess temperature T_p which is varied from 140 to 1260 K above the background mantle temperature.

The sensitivity of the viscosity of the mantle to temperature variations is of particular importance. We use a Newtonian viscosity of the form

$$\eta = \eta_0 \exp\left(A\frac{T_m}{T}\right) \quad (1)$$

where η_0 is defined such that the viscosity equals the reference viscosity η_{ref} at the reference temperature T_{ref} at the bottom of the domain (Table 1). The scaling constant A controls the sensitivity of viscosity to temperature, and T_m is the solidus temperature given by

$$T_m = 1400 + 4.38z - 5.08 \times 10^{-3}z^2 \quad (2)$$

where T_m is in Kelvin and z is depth in km. This solidus fits the melting data of [*Zhang and Herzberg*, 1994] at pressures between 5 and 15 GPa and the linear solidus of [*Takahashi*, 1986] at pressures below 5 GPa. Estimates of A for the mantle range from 30 to 40 [*Borch and Green*, 1989], but these values create extreme viscosity contrasts which are computationally intractable. We varied A from 5 to 15 in order to assess the effect of temperature dependence on the flow. A maximum viscosity of 5×10^{24} Pa s prevents unrealistically high stresses in the lithosphere. Other parameters are given in Table 1 using representative values for the mantles of Earth and Venus.

Results

Isotherms and partial horizontal slices from models with different A and T_p are shown in figure 1 65 Myr after plume initiation. It is apparent from figure 1 that the results of the three-dimensional calculations demonstrate some interesting, new behavior. As T_p is increased keeping A constant (left to right in figure 1) the plume undergoes a transition from the classical, mushroom-shaped axisymmetric plume at low T_p to a multi-armed, non-axisymmetric shape at higher T_p. The same transition occurs as A is increased from 5 to 15 at a constant T_p (bottom to top in figure 1). The axisymmetric mushroom exists under conditions of small A or T_p, while non-axisymmetric forms occur for large A or T_p. Non-axisymmetric as well as axisymmetric, ring-shaped instabilities have been observed in laboratory experiments of plumes [*Griffiths and Campbell*, 1991], but such instabilities have not been investigated numerically except through a mean-field theory calculation [*Yuen and Fleitout*, 1985]. Non-axisymmetric features appear at different scales as the plume interacts with the pre-existing downwellings during spreading and later as the small-scale instabilities form.

Table 1. Model Parameters and Values

Parameter	Symbol	Value
Domain Depth	D	400 km
Grav. Accel.	g	9.8 m s^{-2}
Thermal Expans.	α	3.1×10^{-5} K^{-1}
Thermal Diffusivity	κ	10^{-6} m^2 s^{-1}
Reference Density	ρ_0	3400 kg m^{-3}
Reference Temp.	T_{ref}	1700 K
Surface Temp.	T_{surf}	300 K
Temp. Drop	ΔT	1400 K
Reference Viscosity	η_{ref}	10^{21} Pa s
Rayleigh Number	Ra	9.3×10^4
Whole Mantle Ra	Ra_M	4×10^7

Onset of Instabilities

The temperature at which the transition occurs is higher at lower values of A. This suggests that the non-axisymmetric shapes are due to convective boundary layer instabilities. In a temperature-dependent viscosity fluid, such instabilities are governed by a rheological temperature scale [*Morris and Canright*, 1984; *Davaille and Jaupart*, 1994] given by

$$\Delta T_{rheol} = \left|\frac{\eta}{d\eta/dT}\right|_{T=T_{ref}} \quad (3)$$

From (1), $\Delta T_{rheol} = T_{ref}^2/AT_m$ which is about 50 K for Earth and Venus and about 100 K for the lower pressure Martian mantle. This is the temperature increase that gives an e^{-1} reduction in viscosity and drives convection in an active boundary layer. When the plume arrives beneath the lithosphere, it displaces the marginally stable boundary layer (via mechanical thinning) and a new active layer is produced with a higher effective Ra. This layer is unstable to a smaller scale of convection which acts to reduce the viscosity contrasts in the boundary layer to a factor of 10-20 as observed in fluid tank [*Nataf and Richter*, 1982; *Davaille and Jaupart*, 1994] and numerical [*Christensen*, 1984; *Moresi and Solomatov*, 1995] experiments. When T_p is scaled by ΔT_{rheol}, the transition from axisymmetric to non-axisymmetric plumes is found to occur at a non-dimensional temperature ($T_p/\Delta T_{rheol}$) of about 2, independent of A. We therefore have a robust means of extrapolating the results of the present calculations at low A (15 or less) up to values of A from 30 to 40 relevant to planetary mantles [*Borch and Green*, 1989]. This leads to the important result that mantle plumes will excite non-axisymmetric instabilities for plume temperatures more than a modest 100 K (Venus and Earth) to 200 K (Mars) above the background mantle.

Axisymmetric vs. Three-Dimensional Models

A comparison of the axisymmetric part of the three-dimensional temperature fields with T_p equal to 700 K

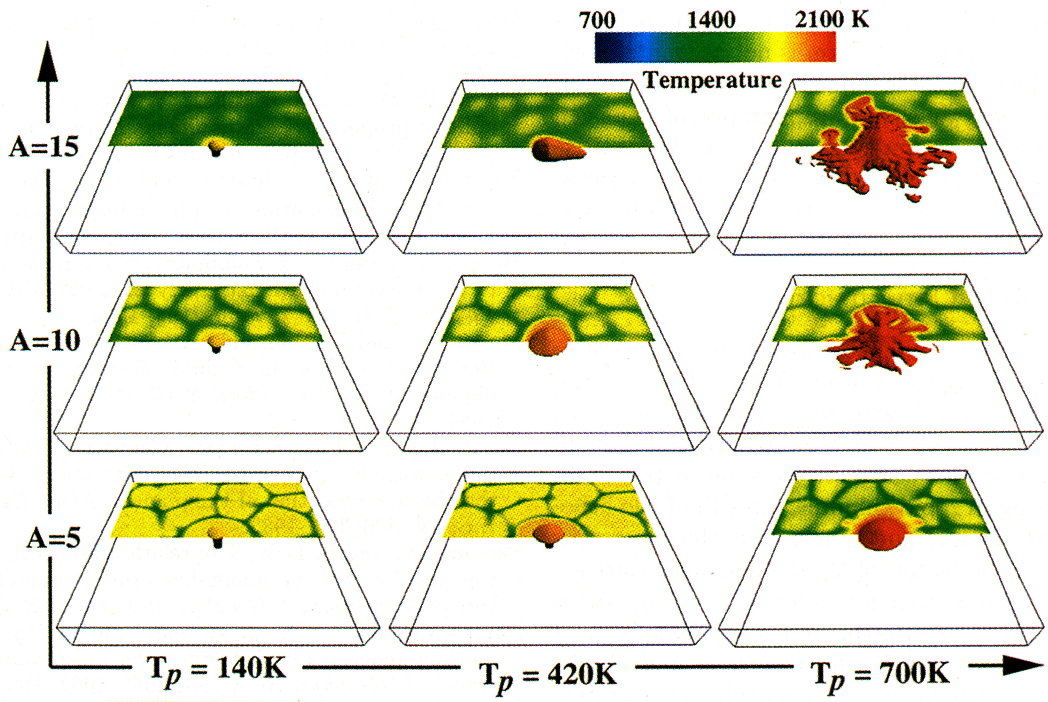

Figure 1. Three-dimensional isotherms at one-half T_p with partial horizontal slices at mid-depth. The axes give the variations of A and T_p. The models are shown 65 Myr after plume initiation

(right column of figure 2) with the results of strictly axisymmetric calculations with identical parameters (left column of figure 2) reveals that the non-axisymmetric modes present for A greater than 10 enhance lithospheric thinning over the axisymmetric case. The enhancement is a factor of three at the highest A modeled and increases with A (scaling roughly linearly with $T_p/\Delta T_{rheol}$), showing that lithospheric thinning is underestimated in axisymmetric models. The efficiency of lithospheric thinning by small-scale flow confirms mean-field results of *Yuen and Fleitout* [1985].

Thinning Rates

The rates of thinning we have obtained in our models are a function of $T_p/\Delta T_{rheol}$ and are a few (1-4) kilometers per million years, resulting in the factor of three thinning in 65 Myr shown in the upper-right frame of figure 2. Thinning rates are higher at higher $T_p/\Delta T_{rheol}$. These rates depend somewhat on the initial thickness of the lithosphere, due to the dependence of ΔT_{rheol} on depth. The lithospheric thickness variations implied by these high thinning rates are comparable to those observed for the Venusian volcanic rises [*Moore and Schubert*, 1997b].

An important question is whether these rates are sufficient to produce thinning of a moving lithosphere such as the majority of Earth's tectonic plates. We have presented results for this case in [*Moore et al.*, 1998] and find a similar scaling for the onset of small-scale instabilities which promote thinning. The stability of the boundary layer to small-scale modes is not particularly enhanced by its large-scale motion over the plume since the plume material is rapidly swept along with the plate, eliminating the relative motion.

Many of the physical processes neglected by our simple model (melting and compositional buoyancy, nonlinear rheology, compressibility and viscous dissipation) is likely to be de-stabilizing to the non-axisymmetric modes investigated here, so that the thinning rates presented here are most likely lower bounds. With a permeable lower boundary, we are not able to determine

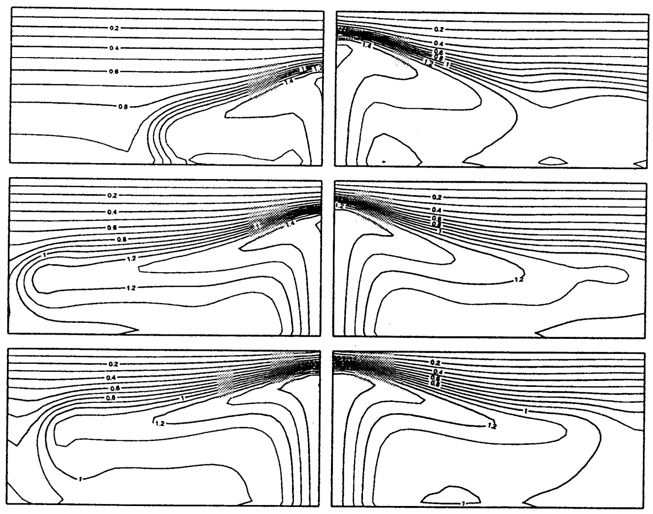

Figure 2. Non-dimensional temperature contours from axisymmetric (left) and azimuthally averaged three-dimensional (right) solutions. T_p is 700 K and A is 15 (top), 10 (middle), and 5 (bottom). The simulation time is 65 Myr after plume initiation. The origins of the plumes are along the center of the figure, and only half the horizontal extent of the domain is shown.

quantitatively reliable topography and geoid anomalies (flow in the mantle below 400 km is not accounted for), but topography from these models is comparable in amplitude to the long-wavelength topography of Venusian volcanic rises (1 to 3 km) and the Cape Verde Rise (2 km) on the African plate. Surface heat flow variations in our models are similar to the 30 to 40% observed at Cape Verde [*Courtney and White*, 1986].

Summary and Conclusions

In summary, we have demonstrated that small-scale instabilities are likely to be promoted by plumes in the mantles of Venus and Earth (under Africa) for plume temperatures about 100 K above the background mantle. These instabilities advectively remove the lithosphere, resulting in rapid thinning rates (half the lithosphere in tens of Myr) and lithospheric thickness variations of up to a factor of 3 or 4, sufficient to account for those observed for volcanic rises on Venus [*Moore and Schubert*, 1997b] and suggested by heat flow anomalies at Cape Verde [*Courtney and White*, 1986]. Plume models calculated in axisymmetric or two-dimensional geometry artificially suppress these instabilities and therefore drastically underestimate thinning rates for Earth-like rheology and plume temperatures.

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