On the transient nature of mantle plumes

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We present new experiments focussing on the transient behaviour of thermal plumes. In a fluid heated from below, plumes develop once the hot thermal boundary layer (TBL) reaches a critical thickness (Howard, 1964). They rise through the fluid owing to their thermal buoyancy and comprise TBL material which empties itself into the plumes. As the TBL becomes exhausted, plumes start disappearing from the bottom up, sometimes even before reaching the upper boundary, depending on the convection intensity. Then, they finally fade away by thermal diffusion. This sequence of events shows that time-dependence is a key-factor when interpreting present-day tomographic images of mantle upwellings. In particular, it could be erroneous to identify the depth of a present-day slow seismic anomaly with the depth of its origin, or to interpret the absence of a long tail as the absence of a plume.

1. Introduction

Our image of what mantle upwellings should look like is still dominated by plumes issued from a steady point source of buoyancy (e.g. Whitehead & Luther, 1975; Olson & Singer, 1985; Campbell & Griffiths, 1990). In a fluid whose viscosity depends strongly on temperature, this set-up generates steady "cavity plumes" with large heads and thin trailing conduits (e.g. Campbell & Griffiths, 1990). In this framework, long-lived hotspot tracks would be produced by the impingement under the lithosphere of plume conduits of 100 km in diameter connecting the lithosphere to the transition zone or to the core-mantle boundary (e.g. Morgan, 1971), and thetraps observed at the beginning of some tracks by the impingement of heads of 1000 km in diameter (e.g. Richards & al, 1989). However, in a fluid heated (even partially) from below like the mantle, plumes develop as thermal boundary layer (TBL) instabilities from the hot bottom boundary and are transient features (e.g. Sparrow & al, 1970). Howard (1964) developed a phenomenological model for plume's formation, and gave scalings for the onset time of thermal instabilities and the boundary heat flux which explains well the experimental data (e.g. Sparrow & al 1970, Davaille & Jaupart, 1993). But the detailed sequence of plume formation and disappearance has not yet been studied in detail, due to experimental limitations.

We therefore present in section 2 new experiments designed to focus on the birth, life and death of a plume issued from a hot TBL. They show that the phenomenological model of Howard (1964) is quantitatively correct,

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and allow to put bounds on plume duration. Those results are then used in section 3 to reinterpret recent images from regional and global tomographic models.

Our image of what mantle upwellings should look like is still dominated by plumes from a steady point source of buoyancy (e.g. Olson & Singer, 1985; Moses & al, 1993). In a fluid whose viscosity depends strongly on temperature, this set-up generates steady "cavity plumes" with large heads and thin trailing conduits. In this framework, the flood basalts observed at the beginning of some tracks would be produced by the impingement of heads, and long-lived hotspot tracks by hot plume conduits connecting the lithosphere to the transition zone or the core-mantle boundary (e.g. Morgan, 1971). Recent tomographic images seem indeed to display such conduits beneath about 10 hotspots. However, no deep slow anomalies or only disconnected ones are reported beneath most hotspots (e.g. Ritsema & Allen 2002, Montelli & al 2004). This may be more coherent with convection in a fluid heated (even partially) from below like the mantle, where plumes develop as transient instabilities of a hot (TBL) thermal boundary layer (e.g. Sparrow & al, 1970; Sleep, 1992; Trompert & Hansen, 1998; Xi & al, 2004). Howard (1964) developed a phenomenological model for plume formation, and gave scalings for the onset time of thermal instabilities and the boundary heat flux which agree well with experimental data (e.g. Sparrow & al, 1970; Manga & Weeraratne, 1999; Parmentier & Sotin 2000). But the detailed sequence of local plume formation and disappearance has not yet been studied, due to experimental and numerical limitations. We present in section 2 new laboratory experiments designed to focus on the birth, life and death of a plume arising from a hot TBL. Simultaneous in situ visualization of isotherms and velocity fields allows us to characterize its morphology and to put bounds on its lifetime. We then use these results in section 3 to reinterpret recent images from regional and global tomographic models.

2. Plumes as thermal boundary layer instabilities

2.1. Necessary conditions for plume generation

The intensity of convection is related to the global Rayleigh number, which compares the driving thermal buoyancy forces to the resisting effects of thermal diffusion and viscous dissipation :

$$Ra(H, \Delta T) = \frac{\alpha\ g\ \Delta T\ H^3}{\nu\ \kappa}, \tag{1}$$

where H is the layer depth, ΔT the temperature difference applied across it, g the gravitational acceleration, α the thermal expansivity, κ the thermal diffusivity and ν the kinematic viscosity. Convective motions occur when Ra exceeds a threshold Ra_c around 1000, the exact value of which depends on the mechanical and thermal boundary conditions (e.g. Chandrasekhar 1961). Regime diagrams have been determined in the limit $\nu/\kappa \gg 1$ relevant to the Earth (e.g. Travis & al, 1990; Trompert & Hansen, 1998; Manga & Weeraratne, 1999). Just above Ra_c, the convective pattern is stationary, first in the form of 2D-rolls, then as 3D-cells, which eventually become time-dependent around $Ra = 10^5$. Above $Ra = 10^6$, well-defined cells no longer exist; instead, heat is transported across the tank by "plumes" which arise through instabilities of the upper cold and lower hot TBL. These plumes either rise vertically and become disconnected from the TBLs (Sparrow & al, 1970; Xi & al, 2004), or become organized in long-lived nearly vertical columns which migrate horizontally and persist until they merge together (Houseman, 1990; Trompert & Hansen, 1998; Parmentier & Sotin, 2000; Xi & al, 2004).

The viscosity of mantle material depends strongly on temperature. In such a fluid, a stagnant layer develops below the upper cold boundary as soon as the viscosity ratio across the cold TBL exceeds about 10. Beneath this lid, flow closely resembles isoviscous convection driven by a reduced temperature difference (e.g. Richter & al, 1983), and hot thermal plumes are generated in conditions similar to the constant-viscosity case (e.g. Trompert & Hansen 1998; Manga & Weeraratne, 1999).

Hence, there are two conditions for the existence of plumes: 1) the existence of an interface from which a TBL can develop, and 2) a Rayleigh number for the fluid layer involved in convection exceeding 10^6. Just above the core-mantle boundary, the D" layer, with a temperature contrast ranging between 1000 and 1500°K, is a good candidate for a hot TBL. TBLs have also been proposed to exist at the 660-km phase transition or in the mid-mantle if the mantle were strongly or even locally stratified (for a review, see Tackley, 2000). Fig.1 shows the convective regime of a mantle layer as a function of its thickness and viscosity. Plumes can develop in the upper mantle (of thickness 660 km) only if its viscosity is lower than 10^{21} Pa.s. For an average mantle viscosity of 10^{22} Pa.s, plumes will be generated only if the mantle layer thickness exceeds 2000 km. Convective motions originating at the CMB and developing over the whole mantle thickness should therefore take the form of plumes.

2.2. Evolution of a hot thermal instability

Experiments were performed with isoviscous fluids (silicone oils or aqueous polymeric solutions) and sugar syrup with temperature-dependent viscosity, for $5\times10^5 < Ra < 5\times10^7$ and Prandtl number $Pr > 10^3$. We developed a new experimental technique using liquid crystals and glass particles ($< 40\mu$m in diameter), to visualize simultaneously the thermal and velocity fields in a planar vertical cross-section of our tank (Davaille & al, 2005).

Fig.2 shows the development of convective instabilities in a layer of sugar syrup initially at uniform temperature T_m, suddenly heated from below at a constant temperature $T_m + \Delta T_m$ for $Ra = 1.7 \times 10^6$. At first, the isotherms (bright lines) remain horizontal and the temperature front moves away from the boundary by conduction (fig.2 a): there is no motion in the fluid. Then the conduction layer (or TBL) contained between the moving front and the outer boundary suddenly becomes unstable (fig.2 b) and breaks up (fig.2 c). It produces a mushroom-shaped plume (fig.2 c, e, g) with a relatively small head (radius less than twice the stem radius; fig.2 c, e), due to the small viscosity contrast with the ambient fluid. The tip of the head ascends initially with a constant velocity (fig.3), and slows down when it nears the upper boundary. The point of maximum vertical velocity is initially within the head (fig.2 d), and recedes into the stem as the head slows down (fig.2 f, h). In this stage, the head is fed by the stem as in a plume from a fixed source (e.g. Olson & Singer, 1985). When the plume reaches the top boundary, it spreads under it, forming a large pond (fig.2 g). During its ascent, the plume also cools by heat diffusion as indicated by the progressive disappearance of the hottest isotherm (40.5°C) from fig.2 c to fig.2 g. Meanwhile, the TBL empties itself into the plume (fig.2 d, f). Once its entire contents have been exhausted, the plume tail starts to disappear from the bottom boundary upward both on the temperature and on the velocity field (fig.2 g, h). At the end, only the pond below the top boundary is visible, and is slowly disappearing as the plume is cooling.

The same time-sequence and morphology are observed for all three fluids used (fig.4). Hence, a TBL instability is an hybrid feature between an isolated hot pocket or "thermal" (e.g. Griffiths, 1986) and a plume issued from a steady point heat source (e.g. Moses & al, 1993).

2.3. Time Scales

The longest stage of plume formation is the conductive TBL building (fig.2, 3). Then, once it has been exhausted, the TBL is locally rebuilt by heat conduction, starting a new cycle. Because TBL instabilities develop from variable perturbations, the cycle duration has a standard deviation of about 10%. Thus although the convection onset is quasi-synchroneous over the whole tank (fig.2), subsequent instabilities are not, and several plumes at different stages of their development are seen at a given time (fig.4). On the other hand, even though there is no coherent large-scale flow, plume formation at a given location is nearly periodic as described above.

Both the TBL cycle duration τ and the plume lifetime τ_p are proportional to the time τ_c at which the TBL becomes unstable. Howard (1964) suggested that this happens when the thickness $\delta = \sqrt{\pi\kappa\tau_c}$ of the TBL is such that the local Rayleigh number $Ra(\delta, \Delta T_m)$ exceeds the critical value Ra_c, which gives:

$$\tau_c = \frac{H^2}{\pi\ \kappa} \left(\frac{Ra_c}{Ra_m}\right)^{2/3}. \qquad (2)$$

Experimental measurements of τ_c agree with eqn.(2) if $Ra_c = 1300 \pm 500$ (Le Bars & Davaille 2004). This value is close to the marginal stability value of 1100 predicted for a linear temperature gradient between rigid and free-slip isothermal boundaries, which are the conditions prevailing across the TBL.

Estimating the recurrence time τ between two TBL instabilities on fig.2 and 3 as the time of plume detachment, we find $t = 550$s, so that the conductive stage occupies 67% (370s) of the whole cycle (fig.3 a). This confirms Sparrow & al (1970) and Manga & Weeraratne (1999)'s resultswherein eqn.(2) fitted cyclicity measurements if $Ra_c = 2050 \pm 300$. Hence, albeit about 30% longer, τ scales as τ_c.

Once a plume has been generated from a TBL instability at $t = \tau_c$, it rises, ponds under the upper boundary and cools by thermal diffusion (fig.2). Since the typical plume size is comparable to the TBL thickness δ, the plume's lifetime is thus also comparable to τ_c. So eqn.(2) gives an upper bound for plume lifetime. For the mantle, eqn.(2) predicts plume lifetimes of 10, 40 and 200 Myr for viscosities 10^{19} Pa.s, 5×10^{20} Pa.s, and 10^{22} Pa.s, respectively.

3. Interpreting tomographic images : Variety of plumes or variety of stages?

In our experiments, the isotherms outline the ascending hot plume as an iso-velocity curve would do on a tomographic map. It is therefore tempting to compare our images to tomographic ones. Since a plume issued from a hot TBL is transient, the classical image of the mantle plume with a long conduit connecting its source in a TBL to the lithosphere is valid at best only during a time which is much shorter than the plume's lifetime (fig.4 a). In some cases it never occurs at all, the plume arriving under the lithosphere being already disconnected from the lower TBL (fig.1 e). Hence, mantle plumes may have remained "elusive" (Ritsema & Allen, 2003) in tomographic images in part because we were looking for an oversimplified cartoon. Since tomographic images only provide instantaneous present-day images of the mantle, plumes might not be so easy to detect in them.

In a recent study, Montelli & al (2004) showed images of long cylinders of slow seismic P-wave velocity anomalies underneath most of the major hotspots. Some of them faded at mid-mantle depths. According to our results, those slow cylinders could be the signature of transient thermal plumes, having originated at the CMB and subsequently detached from it (fig.2 g, 4 b). There would be no need to invoke a mid-mantle thermal boundary layer whose signature is anyway lacking (e.g. Tackley, 2000).

Other tomographic studies show only shallow structures beneath some hotspots (e.g. Ritsema & Allen, 2003). The Azores, for example, are characterized by a strong broad negative seismic velocity anomaly (fig.5) which is clearly visible at 100 km depth, but disappears below 250 km (Silveira & Stutzmann, 2002; Pilidou & al, 2005). At these depths, both temperature, water content and partial melt act to decrease seismic wave velocities. Hence the negative velocity anomaly above 200 km probably corresponds to a combination of the three effects. Since temperature and volatile content also reduce density, the negative seismic anomaly indicates buoyant material, and could be produced by a plume. According to section 2 the existence of a slow pond right under the Azores together with the lack of conduit in tomographic models (fig.5) could be the signature of a dying plume (Silveira & al, 2004).

4. Conclusions

A TBL instability is an hybrid feature between a pure "thermal" and a plume issuing from a steady point heat source. The classical image of a plume with a broad head followed by a long narrower tail is valid at best only during a short period of a plume lifetime. We focussed here on the simplest case of a purely thermal plume, but the same kind of time-dependence is observed in thermo-chemical convection (e.g. Davaille & al, 2003). It may therefore be erroneous: 1) to identify the depth of a present-day slow seismic anomaly with the depth of its origin, and 2) to interpret the absence of a long tail as the absence of a plume. Time-dependence is a key-factor when interpreting present-day tomographic images of mantle upwellings.

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Figure 1. Convective regime developping in a mantle layer as a function of its thickness and viscosity. The pattern is cellular for $Ra \geq Ra_c = 650$, and in plumes for $Ra \geq 10^6$. The grey band shows the average viscosity inferred for the whole mantle (e.g. Ricard & al 1989). The rectangle at the upper left delimits the upper mantle. The calculation has been done with eqn.(1) and $\kappa = 10^{-6}$ m^2/s, $\alpha = 2.10^{-5}$ $^\circ$K^{-1} and $\Delta T = 3000$K.

Figure 2. Thermal boundary layer instabilities in a layer of sugar syrup, initially at 21°C and suddenly heated from below at 53°C ($Ra = 1.7\ 10^6$). a) Negative of the picture taken at $t = 300$ s. The isotherms appear as white lines. The TBL is growing by conduction from the lower boundary. b) $t = 400$ s. The TBL becomes unstable. c) $t = 460$ s. A thermal plume, outlined by the 24.6°C isotherm, rises from the TBL, and the TBL begins to shrink. d) Corresponding velocity field deduced from PIV. The colour background represents the velocity magnitude. e) $t = 500$ s. The plume is well developped and the TBL is emptying itself into it. f) Corresponding velocity field. g) $t = 600$ s. The plume head has reached the upper boundary and begins to spread under it. The TBL has nearly disappeared and the conduit is disconnected from its source. h) Corresponding velocity field.

Figure 3. a) Height (normalized by the tank depth) of the 24.6°C isotherm as a function of time, and b) of the uplift velocity. The plume ascent is rapid compared to the conductive stage. The plume becomes detached from the TBL at $t = 550$ s.

Figure 4. Experiment in constant viscosity fluid (aqueous natrosol solution; $Ra = 2.10^6$). Only the 31.4°C isotherm was used, corresponding to 0.7 ΔT. Three upwellings at different stages of their development can be seen. a) A plume has attained the upper boundary while still being connected to the lower TBL. b) It has become disconnected 20 s later.

Figure 5. West-East (-50°E to 0°) vertical tomographic cross-section below the Azores at 38°N, from a S-wave anisotropic model (Silveira & al 2004).