

Suction mechanism for iron entrainment into the lower mantle

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[1] Perturbations in the Earth's rotation rate at decadal time periods strongly favor the existence of dissipative coupling at the Core–Mantle Boundary (CMB). Here, we explored the plausibility of maintaining a conducting layer on the mantle-side of the CMB, which can couple the outer core and mantle through Lorentz torques. We propose a suction mechanism that maintains a porous medium on the mantle side of the CMB, with the interconnected pore-space partly or entirely filled with liquid iron up to a thickness of ~ 1 km. The suction arises from the deviatoric stresses supported by the mantle-solid in regions of mantle downwelling. Infiltration of liquid iron occurs by percolation, but is inhibited by the rate of viscous dilation of the solid mantle. Our model enables core-mantle material exchange, and maintains a thin conducting layer that has seismic detection potential. Our model is only marginally satisfactory in explaining the inferred CMB coupling. **Citation:** Kanda, R. V. S., and D. J. Stevenson (2006), Suction mechanism for iron entrainment into the lower mantle, *Geophys. Res. Lett.*, 33, L02310, doi:10.1029/2005GL025009.

1. Background

[2] Previous research has shown a link between perturbations in amplitudes of nutations having decadal periods and transfer of momentum between the outer core and mantle. A CMB conducting layer is one mechanism that can facilitate such a transfer [Buffett, 1992; Holme, 1998a; Buffett *et al.*, 2002]. Nutations are components of Earth's rotation that remain after subtracting out precession. Comparing high-accuracy Very Long Baseline Interferometry (VLBI) observations of nutations to current theoretical nutation models suggests that discrepancies of the order of ~ 0.5 mas still exist in predicting the amplitudes of the 18.6 yr and 1 yr out-of-phase (dissipative) nutations [Buffett *et al.*, 2002]. Current theoretical nutation models include the effects of the elasticity of the Earth, its liquid outer core and solid inner core. They also include dissipative effects like mantle anelasticity and ocean-tides. Theoretical modeling indicates that a core mantle boundary (CMB) coupling-torque of $\sim 10^{17}$ Nm can explain a large part of the discrepancy noted above [Holme, 1998a; Buffett *et al.*, 2002]. One dissipative mechanism is viscous coupling - but we can rule that out because of the centimeter sized viscous boundary layer expected for outer core viscosities of 10^{-2} Pa.s.

[3] A more likely coupling mechanism is that of a thin conducting layer just above the CMB, which can couple the

core and mantle through Lorentz torques associated with eddy currents. The conductance of this layer - defined as the product of conductivity and thickness of the conducting layer in some average sense - strongly influences the total Lorentz torque at the CMB [Holme, 1998a, 1998b]. A coupling-torque of $\sim 10^{17}$ Nm, inferred from matching nutation theory to VLBI data, requires that this layer have a conductance of 10^8 S or greater [Buffett, 1992]. Also, given that this layer is of the order of a kilometer or less, it must have metallic conductivities. If we define the mantle geodetically then it must include all materials that move almost rigidly with the mantle (irrespective of composition) and the conducting layer must lie on the mantle side of the CMB. It could nonetheless have formed by upward sedimentation of material from the core, as proposed by Buffett *et al.* [2000]. Their model requires either very small (sub-micron) grain size for the silicate sediments to trap iron (thereby making the permeability very low) or a very high compaction viscosity (thereby preventing the iron from escaping despite assuming high permeability). The model we describe below is an antithesis in either or both respects. Petford *et al.* [2005] invoke shear-enhanced dilation in mantle downwelling regions to drive liquid flow into the lowermost mantle. However, their assumed rheology, parameter space (with strain rates $> 10^{-12}$ s⁻¹, with shear induced flow dominating over buoyancy driven flow; and infiltration times < 1 yr assuming liquid viscosity $\sim 10^{-2}$ Pa.s), and boundary conditions (infinite permeability at the base; impermeable at the top) seem more appropriate for crustal phenomena like emplacement loading [Koenders and Petford, 2000]. They also ignore the effect of liquid fraction on material strength and permeability. Poirier *et al.* [1998] considered liquid iron infiltrating the lower mantle through capillary action alone. They deduced that the liquid couldn't rise beyond a few tens of meters, which is insufficient for entraining enough liquid iron to yield the required conductance. Here we propose a suction mechanism that maintains a mantle-side conducting layer that is stable for durations similar to mantle convection time scales or longer. Also, our mechanism allows much more liquid iron infiltration into the lower mantle compared to capillary action.

2. A Description of the Suction Mechanism

[4] In the absence of convection, the CMB is an equipotential. In the presence of mantle convection, the non-hydrostatic stress field of that flow creates dynamic topography, h (with $h > 0$ denoting an upward deformation), typically estimated to range from hundreds of meters to several kilometers peak to peak [Hide, 1989; Jault and Le Mouél, 1990; Earle and Shearer, 1997]. Equality of normal stress at the CMB implies that

$$p_l(h) = p_s(h) - 2\eta_s \frac{\partial v_y}{\partial y} \quad (1)$$

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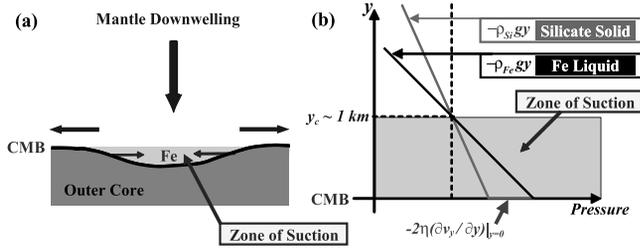


Figure 1. Schematic diagram of suction mechanism: Curves for the hydrostatic pressure heads in the solid and liquid phases intersect at height y_c above the CMB due to the fact that in comparison to the infiltrating liquid, the solid matrix has both a lower pressure gradient in the mantle as well as a lower pressure at the CMB. For $y \leq y_c$, liquid infiltration occurs through grain-boundary processes (since $p_l - p_s \geq 0$ in this region).

where p_l is the pressure in the liquid, p_s is the pressure in the solid, g is the local gravitational acceleration, η_s is the mantle viscosity and v_y is the vertically upward mantle flow (direction y). In the case of a cold, downward and divergent flow (as might occur beneath a subducting slab), $\partial v_y/\partial y < 0$ - so, just above the CMB, the pressure in the solid mantle material (p_s) differs from the pressure of the adjacent core material (p_l) by $p_l - p_s = -2\eta_s \partial v_y/\partial y$, a positive quantity. If there were inter-granular core fluid in hydrostatic pressure equilibrium with the core then this positive pressure difference would drive dilation (negative compaction) of the mantle material, with the porosity filled by core fluid drawn up against gravity. This is what we mean by suction (Figure 1), and we denote the “suction stress” as $\Delta\sigma_{suction} = -2\eta_s \partial v_y/\partial y$ (a positive quantity). This can be treated as a constant on the scale of our interest (~ 1 km) because it is expected to vary on a length scale of mantle convection ($\sim 10^3$ km). We will return to question this assumption later. Now consider a location at a distance y above this deformed CMB and assume that there is an interconnected fluid channel linking that location to the core. In hydrostatic equilibrium, the pressure in the fluid, $-\rho_l g y$, is lower than the pressure at the CMB but the pressure in the solid has changed by only $-\rho_s g y$. Hence, the available pressure difference for driving dilation vanishes at a height,

$$y_c = \frac{\Delta\sigma_{suction}}{\Delta\rho g} \quad (2)$$

where, $\Delta\rho = \rho_l - \rho_s$, a positive quantity, with core-side density, ρ_l , and mantle-side density, ρ_s . The distance y_c is accordingly the equilibrium height to which iron can be sucked from the core, assuming permeability exists and melt fraction is small. Under these conditions, non-hydrostatic pressure contributions are small, and for a CMB location with $h > 0$, $p_l(h) - p_s(h) = -\Delta\rho g h \sim \Delta\rho g y_c$, so that $y_c \sim -h$. The net effect is to raise iron to a radius that is typically less than or comparable to the CMB radius that would exist in the absence of mantle convection. However, iron would be actively squeezed out of regions where $h > 0$. Consequently, our proposed iron infiltration can only occur at roughly half of the CMB.

[5] None of this will happen unless the core fluid can form interconnected pathways in the mantle. This can arise in two ways: Either through interconnection of iron alone or through permeability pathways provided by partial melting of the mantle (as inferred from seismic ultra low velocity zones (ULVZs)). Support for good connectivity of the infiltrating liquid iron comes from high temperature and pressure experiments which show that liquid iron-rich metal can make interconnections even at very small melt-fractions, and that the dihedral angle decreases to below 60° at CMB conditions [Takafuji *et al.*, 2003]. For favorable dihedral angles, it has been shown both theoretically [Stevenson, 1986] and experimentally [Riley *et al.*, 1990] that liquid iron can penetrate an initially “dry” mantle by surface tension driven infiltration. The fact that capillary effects are small [Poirier *et al.*, 1998] is irrelevant since the primary driving force for extensive infiltration is still the suction; capillary effect is needed only to initiate the pathways and this process is fast.

[6] The full complexity of the standard description of percolative flow and compaction [McKenzie, 1984; Scott and Stevenson, 1986] is not needed to understand the essential features of this problem. According to Darcy’s law the vertical flux (volume across unit area per unit time), $w = -[k(f)/\eta_l] \partial [p_l + \rho_l g y]/\partial y$, where $k(f)$ is the permeability, η_l is the liquid (core) viscosity and the y -derivative represents the deviation of the pressure gradient from hydrostatic equilibrium. For very low liquid fractions, vertical force balance gives,

$$\frac{\partial}{\partial y} [\sigma_{yy} + \rho_s g y] = \frac{\partial}{\partial y} [p_s + \Delta\sigma_{suction} + \rho_s g y] = 0 \quad (3)$$

Multiplying (3) by $\{-k(f)/\eta_l\}$, adding it to the Darcy flux above, and re-arranging,

$$w = -\frac{k(f)}{\eta_l} \frac{\partial}{\partial y} \left[\Delta\rho g y + \zeta_s \frac{\partial f}{\partial t} - \Delta\sigma_{suction} \right] = -\frac{k(f)}{\eta_l} \frac{\partial P_{drv}}{\partial y} \quad (4a)$$

where we have used $p_l - p_s = \zeta_s \partial f/\partial t$, which is the definition of bulk or “compaction” viscosity, ζ_s , and $\partial P_{drv}/\partial y$ is the driving pressure gradient. By continuity,

$$\frac{\partial w}{\partial y} = -\frac{\partial f}{\partial t} \quad (4b)$$

There are two limiting regimes for (4a): the simple Darcy flow regime and the compaction-dominated regime. In the simple Darcy flow regime, $\partial P_{drv}/\partial y$ is large, typically of order the gravity term ($\Delta\rho g$) alone. However, we can easily see that this is likely to be an uninteresting regime for us for the following reason: A typical estimate for the permeability is $10^{-3} d^2 f^2$ where d is the grain size; For $\eta_l = 10^{-3}$ Pa.s, $d = 10^{-3}$ m, and $f = 0.1$, this yields a Darcy flux of $\sim 10^{-3}$ m/s, sufficient to fill (or drain) a layer of 1 km thickness in $\sim 10^6$ sec. On the other hand, the timescale to achieve a volumetric strain of ~ 0.1 using a stress (suction) of 10^7 Pa is $\zeta_s/10^8 \sim 10^{12-14}$ sec, for $\zeta_s \sim 10^{20-22}$ Pa.s. This long time scale compared to simple Darcy flow tells us we are very strongly in the compaction-dominated regime. What this means is that (except for

transient responses) $\partial P_{dr}/\partial y$ must be $<10^{-6} \Delta \rho g$ for w to be compatible with the continuity (4b). We can accordingly set $\partial P_{dr}/\partial y = 0$ and immediately integrate to yield

$$\Delta \rho g y + \zeta_s \frac{\partial f}{\partial t} = \Delta \sigma_{suction} \quad (5)$$

where the zero constant of integration is enforced by the CMB condition, $(p_l - p_s)|_{CMB} = (\zeta_s \partial f / \partial t)|_{CMB} = \Delta \sigma_{suction}|_{CMB}$. This can be thought of as “equilibrium” (even though it still predicts a time dependent melt fraction) because it produces an evolution that is restricted to the long time scale ($10^{12} - 10^{14}$ sec) and filters out the transients ($\sim 10^6$ sec). Using the CMB condition along with $(p_l - p_s)|_{yc} = (\zeta_s \partial f / \partial t)|_{yc} = 0$ at the upper boundary, the solution to (5) is,

$$f = \frac{\Delta \sigma_{suction}}{\zeta_s} \left(1 - \frac{y}{y_c}\right) t \quad (6a)$$

(6a) therefore predicts a large melt fraction in the aforementioned geologically short timescale of $\sim 10^{12} - 10^{14}$ sec. We can now estimate the conductance by Archie’s Law [Buffett *et al.*, 2000]:

$$\begin{aligned} C(t) &= \sigma_c \int_0^{y_c} \{f(y, t)\}^n dy = \sigma_c \left(\frac{\Delta \sigma_{suction}}{\zeta_s} t\right)^n \int_0^{y_c} \left(1 - \frac{y}{y_c}\right)^n dy \\ &= \sigma_c \left(\frac{\Delta \sigma_{suction}}{\zeta_s} t\right)^n \frac{y_c}{(n+1)} \end{aligned} \quad (6b)$$

Here, $n = 1-2$. If we assume fully-connected liquid fraction ($n = 1$), we obtain a conductance, C , of $\sim 10^8$ S in $\sim 10^{14}$ s (10 Ma), with $\Delta \sigma_{suction} = -2\eta_s \{\partial v_y / \partial y\}|_{y=0} \sim 10^7$ Pa, $\zeta_s \sim 10^{22}$ Pa.s = η_s , $\sigma_c = 10^6$ S/m, and $y_c \sim 1000$ m.

[7] There is an obvious shortcoming of this analysis: It predicts that the melt fraction grows without bound. Evidently, the mechanism will saturate once the melt fraction becomes sufficiently high that the deviatoric stress responsible for the suction is diminished. This is not a fundamental objection to the mechanism, unless one thought that this weakening of the suction were so severe as to limit the infiltration to low melt fractions. To assess this (but still assuming $f \ll 1$ so that $(1 - f) \sim 1$ in the full theory of compaction and Darcy flow), we need to introduce an f dependence to $\Delta \sigma_{suction}$. In final steady state (infinite time), (5) is replaced by $\Delta \rho g y = \Delta \sigma_{suction}(f) = \Delta \rho g y_c \varphi(f)$, where $\varphi(f)$ describes the weakening of the medium (or modification of the mantle flow field). We expect $\varphi(0) = 1$ and $\varphi(f^*) = 0$, where f^* is the critical melt fraction for disaggregation (catastrophic drop of the matrix viscosity). Plausibly $f^* \sim 0.25$ to 0.3 [Scott and Kohlstedt, 2004]. Although there is no widely accepted functional form for $\varphi(f)$, Xu *et al.* [2004] suggest a strong reduction in viscosity with melt fraction. For example, if one assumed $\varphi(f) = (1 - ff^*)^n$ with $n \sim 2$ to 4 , then the infinite time solution, $f = f^*[1 - (y/y_c)^{1/n}]$. This predicts a fairly low melt fraction except near the CMB where it is guaranteed to reach the highest possible values. Notice that this melt weakening also applies to the compaction viscosity (though not necessarily in exactly the same functional form) with the result that the infiltration may

actually be accelerated relative to (6a) above. However, this has no bearing on the more important issue of the final steady state. If we neglect this complication, the time evolution equation (5) takes the form

$$\frac{\partial f}{\partial t} = \varphi(f) - \frac{y}{y_c} \quad (7)$$

where \tilde{t} is a dimensionless time, measured in units of $\zeta_s / \Delta \sigma_{suction}|_{f=0}$. For the functional form introduced above and illustrative case $n = 2$, the corresponding solution (with $f = 0$ at $t = 0$) is

$$\frac{f}{f^*} = \frac{(1 - x^2) \left\{1 - \exp\left(-\frac{2x}{f^*}\right)\right\}}{\left\{(1 + x) - (1 - x) \exp\left(-\frac{2x}{f^*}\right)\right\}} \quad (8)$$

where $x = \sqrt{(y/y_c)}$. This solution agrees with the infinite time solution and also agrees with our early time solution (6a) for $2xt/f^* \ll 1$. This example should not be viewed as a prediction because it uses a particularly simple functional form for the weakening but it serves to illustrate that the approach to equilibrium is a smooth relaxation behavior.

[8] The full equations for percolative flow with compaction are known to admit wave-like solutions in space and time [Richter and McKenzie, 1984; Scott and Stevenson, 1984, 1986]. Hence, numerical solutions with initial conditions that are out of equilibrium, (6), produce fluctuating melt fraction. This has no bearing on the final steady state or the long term evolution at the CMB, where the suction would increase gradually from zero to its peak value over the lower mantle convection timescale ($\sim 10^8$ years). In such a realistic “adiabatic” evolution, the system can be expected to track the equilibrium solution closely at all times due to the long timescale.

[9] The remaining concern would be the evolution beginning from very small f : clearly, there is a melt fraction so small ($f \sim 10^{-4}$ or 10^{-3}) that the “equilibrium” described above does not apply. It is possible to derive approximate solutions for this situation. However, the time spent in that condition is negligibly small and therefore not interesting.

3. Discussion and Conclusions

[10] Using the infinite time solution for f , we estimate that the conductance for a CMB-wide conducting layer is $\geq 10^8$ S – which is consistent with the value inferred from nutation data. However, the layer obtained by using the suction model does not cover the entire CMB, and may be too thin to yield high enough conductance. So, our model is only marginally satisfactory in explaining the discrepancies between observed and calculated nutations. The suction mechanism may still be active, however, over parts of the CMB. As mentioned above, the suction and sedimentation models are antithetical in the sense that the conditions favorable to one are unfavorable to the other. The suction mechanism raises liquid iron into the topographic depressions of the CMB (relative to the equipotential surface). This mechanism is favored at “low” viscosities ($10^{20} - 10^{22}$ Pa.s). In contrast, buoyancy driven upward sedimentation at the top of the core is plausible only for large

compaction viscosities ($>10^{24}$ Pa.s, assuming permeabilities arising from a realistic grain size of ~ 1 mm, and considering the effect of liquid fraction on sediment viscosity). Also, buoyant sediments are likely to collect in topographic elevations of the CMB. Thus, these two models result in different locations for the conducting layer, the effect of which may be seismically tested due to the fact that they result in different CMB impedance contrasts. For the suction model, the contrast is between a silicate matrix containing $\leq 10\%$ liquid iron versus liquid iron; for the sedimentation model it is between mantle silicate versus a core sediment matrix containing $\leq 5\%$ liquid iron. From an analysis of post-cursors to ScP phases (S-waves that are converted to P-waves upon reflection from the CMB), Rost and Revenaugh [2001] infer the existence of a thin layer at the CMB - below New Caledonia - having a thickness 1 km, S-wave velocities < 1 km/s, and density 10–40% below that for the outermost core. They interpreted this layer to be on the core-side of the CMB, justifying their results based on Poirier *et al.* [1998] and Buffett *et al.* [2000]. However, a mantle-side suction layer with a similar density reduction (for $f \sim 0.1$), located within a CMB topographic trough below the subducting Tonga slab, may still yield post-cursors, but with slightly different amplitudes. So, a systematic study of amplitudes and waveforms of core-reflected and core-diffracted phases, carried out for regions underneath multiple subduction zones, may be able to discriminate between these two models. There is also an important geochemical implication of the suction model if mantle convection does penetrate all the way to the CMB: that the liquid iron infiltrating the mantle could alter it through chemical reactions [Knittle and Jeanloz, 1991; Song and Ahrens, 1994; Dubrovinsky *et al.*, 2003] thus equilibrating at least the bottom few kilometers of the mantle with the outer core. This equilibration may affect the isotopic composition of deep mantle plumes, and hence, of ocean island basalts.

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