WHY EXISTING TERRESTRIAL PLANET THERMAL HISTORY CALCULATIONS SHOULD NOT BE BELIEVED (AND WHAT TO DO ABOUT IT); David J. Stevenson and Seth S. Bittker, 170–25, Caltech, Pasadena, CA 91125.

Introduction. Beginning in the late 1970's, many workers have used simple parameterized convection recipes for analyzing the thermal history of a terrestrial planet [1–3]. These calculations assume that thermally derived density differences are the only important contribution to convective instability. However, all planets undergo irreversible differentiation and the resulting density effects are locally far larger than thermal effects. This is not a novel observation, but the past response has usually been that the total gravitational energy release which accompanies irreversible differentiation is smaller than that associated with thermal convection and hence energetically less important. Although the latter statement is true (excluding core formation which is believed to be completed very early in all cases [4]), it does not get to the heart of the issue, which is this: This correct parameterization of convection requires a *boundary layer* analysis and it is precisely in this layer where compositional effects can be large because of partial melting. Since the residue left from partial melting is generally less dense than the undepleted mantle below (mainly because the melt is relatively Fe-rich), volcanism can *stabilize* the convective system and even shut the convection off, at least temporarily. This is an especially serious effect on one-plate planets because the lithosphere is not fully participating in the return flow.

Background and Input. Compositional effects associated with basaltic production play a role in the subductability of the Earth's oceanic lithosphere [5], the dynamics of matrix flow and melt migration beneath mid-ocean ridges [6] and in the mantle wedge above subduction zones [7], and in the global contraction or expansion of a planet [8]. We can identify the following regimes of interest:

I. The convective temperature profile (conductive lid and adiabatic interior) never intercepts the mantle solidus. Conventional parameterized recipes will then apply; however, this is an uninteresting case, even in smaller planets, since one plate planets usually "run hot" (because one plate planets are less efficient in eliminating heat) and do not lie in this regime, except when all the basaltic component has been flushed out.

II. Adiabat extends into the supersolidus region, though only to a small extent and not so as to deplete all basalt from the material in that region. Here, we can modify the "local" Rayleigh number and devise the following computational scheme, analogous to those used in the past for purely thermal buoyancy (e.g., ref. 3). We set $F = k\Delta T/\delta$, $Ra \equiv g(\alpha\Delta T - \beta f)\delta^3/\nu(T - \Delta T/2)\kappa$, $d(Ra)/d\delta = 0$, and $Ra = 10^3$ at this extremum. Parameters have their usual meanings (F = heat flow, k = thermal conductivity, g = gravity, α = coefficient of thermal expansion, δ = boundary layer thickness, ν = temperature-dependent viscosity, κ = thermal diffusivity, ΔT = temperature drop across the boundary layer) and the crucial change is the βf term representing the density effect arising from a degree of melting f, averaged within the boundary layer. The value of β is a material property that will vary among planets (see below), and allowance must also be made for the *fertility*: the extent to which the mantle still has the "primordial" basaltic (i.e., pyroxene-rich) component.

III. Large excursion of the adiabat into the supersolidus region. In this case, a secondary convection pattern can initiate within the zone of partial melting since all of this zone is denuded of basalt but has an unstable temperature gradient. This convection will *not* penetrate into the deeper mantle, which is intrinsically more dense (more iron-rich). The analysis so far has, however, ignored phase boundaries which will surely complicate these considerations.

Applications. We have done a detailed analysis of regime II, and made some analysis of the implications for Mars. In the Figure, we show total heat flux vs. mantle temperature for a variety of values of β. (Ignore the flat curve at $T < 1600$ K which corresponds to pure conduction.) The smooth upward curvature at small β corresponds to the usual expectation

that heat flow increases dramatically with increasing "mantle temperature" (here defined as the zero pressure extrapolation of the mantle adiabat). But at sufficiently large β (≥ 0.04), we find a turnover in the curves, reflecting the increasing stabilization of convection. What is a reasonable value for β? On Mars, where most estimates of mantle iron content are high, the implication of the work of Finnerty *et al.* [9], for example, is $\beta \approx 0.1$, very large and easily enough to have an *enormous* effect on the convection. The result is that the mantle heats up, enters regime III, and begins to form convective *layers*. The value of β on Earth, Venus, and possibly Mercury is smaller by a factor of about 2, but still big enough to be very important. (Remember, however, that the model is not meant to apply to Earth.)

Implications. Although detailed models of the resulting thermal evolution have not been completed (they are much harder than previous simple models), it is clear that they will be markedly different from previous work. Here are the main differences: (1) A tendency for the interior to *heat up* at depth even in early history, because volcanism stifles convection. This is in contrast to the trend toward monotonic cooling models and volcanism histories [e.g., 10]. (2) A tendency for episodes of volcanism, separated by long periods of time associated with the merging of convecting layers operating in regime III. (3) Delayed heat loss and the likelihood of continued volcanism on Mars.

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