BACKGROUND HEATFLOW ON HOTSPOT PLANETS: IO AND VENUS

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Abstract. On planets where most of the heat is transported to the surface by igneous activity (extrusive volcanism or near-surface intrusions), the surface heatflow at localities well away from regions of current igneous activity need not be even approximately the conductive heatflow through the entire lithosphere but may instead be dominated by the residual heat leaking out from the last igneous event in that locality. On Io, it is likely that \((\kappa \tau)^{1/2} \ll d_L\) (lithosphere thickness \((\kappa = \text{thermal diffusivity}, \tau = \text{typical time between "resurfacing" events})\) and the background heatflow may be very large, comparable or even larger than the current observational heatflow, which is associated with the hotspots alone. This upward revision of Io’s heatflow is compatible with observations and with recent indications of a non–steady tidal and thermal evolution. On Venus, \((\kappa \tau)^{1/2}\) is probably comparable to the lithosphere thickness and the resulting upward revision of heatflow may be only marginally significant, unless magmatic activity is enormously greater than on Earth.

Introduction

On a solid planet, the simplest view of the surface heatflow is expressed by the equation

\[
F = \frac{k\Delta T}{d_T}
\]

where \(F\) is the heat flux, \(k\) is the thermal conductivity, and \(\Delta T\) is the temperature drop across the thermal lithosphere of thickness \(d_T\). This is approximately correct on Earth, despite the delivery of some heat by volcanism and hydrothermal circulation. Lithospheric radiogenic heating changes the result by at most a factor of two. It is believed to also apply for relatively low heatflow planets such as Mercury, Moon, and Mars. It has become increasingly apparent, however, that it need not apply on high heatflow bodies where hotspot volcanism dominates. Io is a striking example of a body where the topography is substantially, implying a thick lithosphere [Schaber, 1982; Nash et al., 1986; Carr, 1986] and the heatflow is greater than about 1400 erg cm\(^{-2}\) s\(^{-1}\) [Johnson et al., 1984], clearly incompatible with equation (1). In a deceptively simple but important paper, O’Reilly and Davies [1981] pointed out that Io delivers its heat through heat pipes to the surface, effectively decoupling the thickness of the lithosphere from the magnitude of the heat flow. A similar though less extreme application to Venus has been proposed by Turcotte [1988].

Although there is increasing acceptance of the failure of equation (1) for hotspot planets, there is still the common assumption (sometimes explicit, often implicit) that the heatflow is expressed by the equation

\[
\frac{\partial T}{\partial t} = \frac{\kappa}{\rho c} \frac{\partial^2 T}{\partial z^2}
\]

(2)

for which the solution is [Landau and Lifshitz, 1959]

\[
T(z,t) = \Delta T \left\{ \text{erf} \left[ \left( x - d / \sqrt{2\kappa t} \right) \right] - \text{erf} \left[ \left( x + d / \sqrt{2\kappa t} \right) \right] \right\}
\]

(3)

The heat flux corresponding to this is

\[
F = \frac{k\Delta T}{\sqrt{\kappa t}} \left( 1 - e^{-d^2/4\kappa t} \right)
\]

(4)

As one would intuitively expect, \(F \propto t^{-1/2}\) (like a cooling half space) for \(t \leq d^2/4\kappa\) and \(F \propto t^{-3/2}\) once the entire interior of the flow is losing heat to the surface and below.
This exact solution does not quite suit our purpose because it is a single event. Consider a model in which resurfacing occurs by emplacing a layer of thickness \( d \) at times \( t, 2t, 3t, \ldots \). We carried out a number of numerical solutions of the diffusion equation under these circumstances. For \( d^2 \ll 4\pi \kappa t \) we found, not surprisingly, that little memory is retained of previous flows and the behavior is well described by eq. (4) with \( t \) measured relative to the time of last emplacement. If \( d^2 \gg 4\pi \kappa t \) then the flows do not cool internally before the next flow is emplaced. In this limit, no lithosphere can exist. In fact, this is not a meaningful limit for our purposes (it might, however, describe the development of a magma ocean).

The most difficult case is the regime where \( d^2/\kappa t \) is not enormously different from unity. This might seem like a special (hence implausible) regime, but it may even be a preferred regime if the eruption process proceeds once the overlying layer has cooled enough to allow fracture. In this regime, an approximate solution can be obtained, by allowing for a fraction \( \epsilon \) of the heat to be retained from each resurfacing event. This suggests an expression for the heat flux of the form

\[
F \simeq \frac{k\Delta T}{\sqrt{\pi \kappa t}} \left[ 1 - (1 - \epsilon)e^{-d^2/4\pi \kappa t} \right]
\]

where the value of \( \epsilon \) was estimated from numerical simulations such as that shown in Figure 1, and is given roughly by

\[
\epsilon \simeq 0.2 \frac{\kappa}{\kappa t} \quad \tau \gg d^2/\kappa
\]

\[
\epsilon \simeq 1 - 0.8 \left( \frac{\kappa}{d^2} \right)^{1/2} \quad \tau \ll d^2/\kappa
\]

By conservation of energy, the time averaged heat flux at any locality is

\[
\bar{F} \simeq \rho C_p \Delta T u(1 - \epsilon)
\]

\[
u \equiv d/\tau
\]

Temperature vs Depth

![Figure 1](image1.png)

Figure 1. A sequence of numerically computed temperature profiles at successive times \( t, 2t, 3t, \ldots \) just prior to a resurfacing event, for the particular case \( d^2 = \kappa t \) and an equilibrium surface temperature appropriate to Io. In each case, zero on the horizontal axis refers to the actual surface, so the residual (unescaped) heat is trapped deeper down after each new resurfacing.

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In the general case, where the value of \( d^2/\kappa t \) is not enormously different from unity. This might seem like a special (hence implausible) regime, but it may even be a preferred regime if the eruption process proceeds once the overlying layer has cooled enough to allow fracture. In this regime, an approximate solution can be obtained, by allowing for a fraction \( \epsilon \) of the heat to be retained from each resurfacing event. This suggests an expression for the heat flux of the form

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\[ Pe = \frac{d^2}{2\kappa} = \frac{ud}{2\kappa} \] (11)

which can be best interpreted as the fractional time of high heatflow (i.e., the ratio of thermal diffusion time for the resurfacing layer to the recurrence time of volcanic events at any given locality). The factor of two in the definition improves this convenient interpretation. The simple example above (eq. (10)) corresponds to the maximal value, \( Pe \approx 1 \). As \( Pe \) decreases, a larger fraction of the heat output occurs from regions with higher heatflows, but there is always a substantial low heatflow component even at low \( Pe \).

**Application to Io**

The anomalous infrared emission properties of Io have long been known but were not correctly interpreted until after the discovery of volcanism during the Voyager 1 flyby. The analysis of both ground-based and Voyager data has led to the unequivocal identification of an internal heatflow of at least 1400 erg cm\(^{-2}\) s\(^{-1}\) and possibly more [Matson et al., 1981; Johnson et al., 1984; Nash et al., 1986; Johnson et al., 1988]. The identification is unequivocal not so much because of the magnitude of this heatflow (which is only about 10% of the absorbed insololation) but because it is associated with temperatures (ranging from 200 K upward) that are high compared with typical surface temperatures. In other words, the "spectral contrast" (the steep rise of brightness temperature with decreasing wavelength) is an essential part of identifying this heatflow. In the context of the models presented above, we must ask: What is the spectral contrast of the background heatflow? The brightness temperature at each wavelength can be evaluated from

\[ B(\lambda, T_b) = \int_{0}^{\infty} f(T) B(\lambda, T) dT \] (12)

where \( B(\lambda, T) \) is the blackbody emission (Planck function) at wavelength \( \lambda \), \( T_b \) is the brightness temperature, and \( f(T) \) is the probability that a surface element of Io lies in the temperature range between \( T \) and \( T + dT \):

\[ f(T) = -\frac{P(t) dF/dt}{dF/dt} \] (13)

with the heatflux \( F \) given by equation (5). We also express

\[ F = \sigma(T^4 - T_o^4) \] (14)

where \( \sigma \) is the Stefan–Boltzmann constant and \( T_o \) is the insolation temperature. Our purpose here is not to model Io specifically (since that requires detailed assumptions about surface properties) but to show how the spectral contrast varies with total heatflow and Piclet number. This is illustrated in Figure 3, for the choice \( T_o = 120 \) K and two background mean heatflows (1500 and 3000 erg cm\(^{-2}\) s\(^{-1}\)). As these calculations show, a large background heatflow is possible without substantial spectral contrast, provided \( Pe \) is quite large. These heatflows correspond to mean resurfacing rates of up to a few cm/yr [Johnson and Soderblom, 1982; Carr, 1986] and imply that the interior of Io is only partially molten [Webb and Stevenson, 1987].

It is difficult to make independent estimates of plausible values for \( Pe \), the ratio of cooling time for a resurfacing event to the time between resurfacing events, especially when allowance is made for near surface intrusive events. Certainly, \( Pe \) is at least as large as the fractional surface of Io that is currently active, assuming that Io is in a typical state at present. (This is a debatable assumption, but probably not wrong by more than a factor of a few; see below.) This suggests \( Pe \gtrsim 10^{-2} \). However, a substantially larger value would seem plausible. For example, a volcanic province that erupts every 10\(^4\) years and lays down three hundred meters of material at each eruption would have \( Pe \approx 0.3, \text{assuming } K \approx 10^{-2} \text{ cm}^2 \text{s}^{-1}. \) If the actual eruption (i.e., period of activity) takes \( \sim 10^4 \) years and all parts of the surface of Io are equally suitable for eruptions then the "active fraction" of the surface would then be \( \sim 10^{-2} \), much less than the value of \( Pe \). Figure 3 then implies that the spectral contrast is low and the globally averaged background heatflow could be large; comparable or even larger than the hotspot component.

Recent analysis of the coupled thermal and orbital evolution of Io are consistent with this possibility. If the total average heat output were less than \( \sim 2000 \text{ erg cm}^2 \text{s}^{-1} \) then it could be argued (e.g., Schubert et al. [1986]) that Io is in steady state, with its orbit steadily expanding at a rate given by Jupiter's tidal Q and its orbital eccentricity determined by the ratio of Io's Q to that of Jupiter [Yoder and Peale, 1981]. However, data on the expansion of Io's orbit suggest a much slower expansion than expected for the steady state model [Lieske, 1987; Greenberg, 1987] and accordingly allow for a higher heatflow, similar to (but not necessarily exactly compatible with) the non–steady state model of Ojakangas and Stevenson [1986]. The higher heatflow arises if the eccentricity of Io's orbit is currently decreasing. A global heat output of 3000–4000 erg cm\(^{-2}\) s\(^{-1}\) is possible. These arguments are only suggestive and the true test is observational.

**Application to Venus**

The current uncertain state of the interpretation of Venus geology [Basilevsky and Head, 1988] renders any statements about Venus conditional at best. If the total
maggmatic activity of Venus is not much different from Earth, as several have argued [Solomon and Head, 1982; Morgan and Phillips, 1983; Phillips and Malin, 1984], then the considerations developed above have marginal relevance. Specifically, consider a recurrence interval of 10^8 years for magmatic activity, with 10 km of material (basaltic crust or gabbroic intrusion) "laid down." This corresponds to P_e ~ 0.03 and ≤ 10% enhancement of the background heatflow. On the basis of crater counts [Barsukov et al., 1986; Grimm and Solomon, 1987], this recurrence interval is about the smallest conceivable. However, it must be stressed that intrusive activity can come close to achieving much of the same effect (in terms of heatflow) as surface lava flows, yet have a less direct impact on geomorphology, so the situation is not fully determined from the radar data. Turcotte [1988] has suggested that the magmatic activity could be two orders of magnitude higher on Venus than on Earth. Under these circumstances, the "background" heatflow will have comparable contributions from conduction through the lithosphere and cooling of previous igneous events, if the lithosphere thickness d_T ~ 100 km.

Concluding Comments

The main message of this paper is a rather obvious one, but one that seems to have been insufficiently appreciated: On planets where volcanism dominates the heatflow, there is no simple relationship between lithospheric thickness and heatflow, either locally or globally, even away from regions of current volcanic activity. There is no basis, either observationally or theoretically, to advocate the minimalist view that the Io heatflow is as low as the hotspot component alone would suggest. Our developing views of Io and (to a lesser extent) Venus will require a better understanding of how volcanism operates on these bodies.

Acknowledgements. Discussions with T. V. Johnson and D. L. Matson, and comments from a reviewer were helpful. Contribution number 4657 from the Division of Geological and Planetary Sciences, California Institute of Technology, Pasadena, California 91125. This work is supported by the NASA planetary geophysics program, grant NAGW-185.

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(Received June 20, 1988; revised October 24, 1988; accepted October 26, 1988.)