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THE ROLE OF NONUNIFORM INTERNAL HEATING IN TRITON'S ENERGY BUDGET;
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Introduction Triton's large heliocentric distance and high albedo, combined with its unusually large silicate mass fraction, make internal heating more important in its energy budget than in that of any other icy satellite. Brown *et al.* [1] have recently estimated that the average radiogenic heat flux F_0 (which is probably between 3.3 and $6.6\,\mathrm{mW\,m^{-2}}$ depending on core size and composition) may equal 5% to 20% of the average absorbed insolation. On a global scale, this additional energy input appreciably increases the thermal emissivity required to be consistent with the observed surface temperature [2]. Brown *et al.* [1] also speculated that spatial variations of the internal heat flux may change the local sublimation-deposition balance enough to lead to observable modifications of the distribution of volatiles on Triton's surface. In this abstract we attempt to estimate the magnitude of internal heat-flux variations due to the insulating effect of the polar caps, to mantle convection, and to cryovolcanism; we evaluate the importance of these variations in modifying the volatile distribution.

Thermal Structure Our model of the thermal and convective structure of Triton's shallow interior, required for the calculations presented below, is as follows. We assume Triton is fully differentiated. We use the parameterized convection model applied by Kirk and Stevenson [3] to Ganymede, but we specify the heat flux (3.3 or $6.6\,\mathrm{mW\,m^{-2}}$ [1]) and solve for the temperature at the base of the ice layer, rather than the reverse. Our calculation includes the strong temperature dependence of the thermal conductivity of ice [4]. We find that the temperature at the base of the thermal lithosphere is between 170 and 200 K, depending on the estimate of ice viscosity used. The corresponding depth is roughly 300 or 150 km, depending on the heat flux. The total thickness of the ice layer is probably close to 350 km, with the result that the layer is stable against convection for $F_0 = 3.3\,\mathrm{mW\,m^{-2}}$ and a melting-point viscosity $\eta_m \geq 2 \times 10^{14}\,\mathrm{Pa\,s}$.

Insulating Polar Caps The thermal conductivity of solid N_2 [5] is several hundred times less than that of H_2O ice [4] at the surface temperature of Triton; a polar cap of modest thickness may therefore be able to significantly modify the latitudinal distribution of heat flow. We have constructed a simple analytic model to test this effect, assuming linear conduction in a spherical shell, constant temperature on the inner boundary (the top of the convecting zone), and a mixed boundary condition on the outside representing the effects of the insulating cap. For simplicity, we chose a cap thickness varying as $\sin^2$(latitude), making solution for the temperature field as a sum of Legendre polynomials relatively straightforward. We find that for a N_2 layer 1 km deep at the poles, the equatorial heat flux is enhanced by 20% and 35% for the lower and upper limits on heat flux, respectively (Fig. 1). The corresponding flux decreases at the pole are roughly twice as large, leading to equator-to-pole flux ratios of 2 : 1 and 4 : 1 in the two cases. The temperature beneath the cap is elevated by roughly 9 K at the pole.

Effect on Frost Stability Models of the current energy balance on Triton (excluding internal heat but including latitudinal variation of albedo) predict the deposition of frost northward of 15° latitude [6]; the time-dependent model of Spencer [7] predicts that seasonal frost deposits currently extend even farther south. Can we account for the absence of obvious bright frost deposits in Triton's northern hemisphere by including the concentration of internal heat toward the equator in the energy balance? The answer would appear to be no. We have performed a stability analysis similar to that of Stansberry *et al.* [6], comparing the diurnally averaged insolation absorbed by a patch with albedo $A_{local} = 0.6$, appropriate to the northern hemisphere, with the global average absorption for $A = 0.8$. Adding the spatially varying internal heat flux calculated above for $F_0 = 6.6\,\mathrm{mW\,m^{-2}}$ shifts the current latitude of equilibrium by only 0°.5. The shifts for $A = 0.9$ [1] or $F_0 = 3.3\,\mathrm{mW\,m^{-2}}$ are even smaller. For $A_{local} = A$ (*cf.* [7]), the northward shift is larger, but still less than 2°.5.

Somewhat less can be said with certainty about the effect of internal heat on the long-term stability of the polar caps. The seasonally averaged insolation varies much less strongly with latitude than the current diurnally averaged insolation, and proportionately larger shifts of the latitude of stability would therefore be expected. Unfortunately, the redistributed internal heat flux for our model cap shape passes through its global mean value almost precisely at the latitude of where seasonally averaged sublimation and deposition balance on a global frost layer. This coincidence, which limits the calculated latitude shift to less than 0°.1, would not occur if more general cap shapes were used in the heatflow model or if an unfrosted equatorial zone were excluded from the energy budget. In any event, the internal heat flux at the center of the polar caps is always reduced relative to that at their edges and beyond. This redistribution of heat would thus act to hasten—perhaps significantly—the retreat of the permanent caps toward the poles predicted as a consequence of the gradient in insolation [7].

Mantle Convection We turn now to processes capable of producing more localized enhancements of heat flux and thus perhaps able to modify the pattern of frost deposition. One candidate is mantle convection: upward heat flux at the top of the mantle is concentrated over zones of upwelling. We can estimate the extent to which variations in the flux will be attenuated across the lithosphere by a conductive model with fixed temperature on the outside of a spherical shell and with flux varying as a spherical harmonic on the interior. The attenuation factor is approximately $\frac{1}{2}(R_m/R)^{-(l+2)}$, where R and R_m are the radii of the planet and the top of the mantle, respectively, and l is the degree of the spherical harmonic. The lowest degree, hence least attenuated, spherical harmonic component of the mantle heat flux will have l equal to the number of convective-cell pairs that fit around the circumference of the planet (for roll convection; for equant convection cells, l will be roughly twice as large, leading to much greater attenuation). Using the results of the parameterized convection calculation above, we have, for $F_0 = 6.6\,\mathrm{mW\,m^{-2}}$, $R_m/R \simeq 0.89$, $l \simeq 15$, and an attenuation factor of roughly 4. The attenuated flux is sufficient to shift the latitude of frost equilibrium by $\pm 2°$ if the amplitude of the flux variation at the mantle is equal to the mean flux. These results are extremely sensitive to the magnitude of F_0; decreasing it to $3.3\,\mathrm{mW\,m^{-2}}$ both decreases R_m/R and increases l, resulting in attenuation factors of 10^3–10^4.

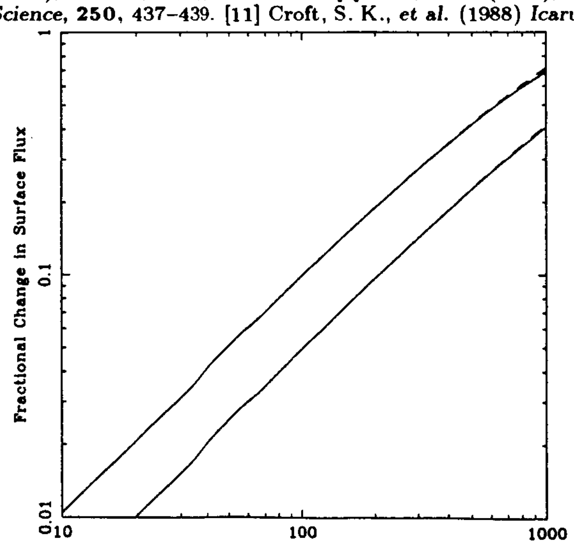
Cryovolcanism The evidence in Voyager images for multiple styles of cryovolcanic eruption on Triton [8] raises the prospect that heat released by cryomagmas may locally contribute to the sublimation of surface frosts. Unlike the mechanisms discussed above, migration of "hot" material toward or to the surface could result in heat fluxes that exceed the global average by a large factor. On the other hand, this enhancement would be transient and hence might not be observed at a given time. We discuss the possible effects of two very different types of cryovolcanic activity.

Fitting the results of viscous spreading models [9] to the measured dimensions of the linear ridges common in Triton's cantaloupe terrain (and extending into the south polar cap) indicates that the erupted material could be subsolidus N_2 or CH_4 or, more plausibly, ammonia-water slush or glass. Independently of the assumed composition, a conduit width on the order of 400 m may be inferred from the flow models and the thickness of the ridges. We have therefore modeled the thermal effect of a dike of this width and initial temperature equal to the NH_3-H_2O eutectic value of 173 K, intruded into a half-space at 38 K. In this linear-conduction model we have used the conductivity of H_2O ice at 38 K; the results are only moderately sensitive to this assumption. We find that the vertical heat flux is enhanced by $3.3\,\mathrm{mW\,m^{-2}}$ or more in a zone extending some 17 km on either side of the dike. We speculate that the narrow swath apparently cleared through the polar cap by a linear ridge at latitude 10°–15° S, longitude 345°–0° may result in part from heating by the surface flow and its conduit. (The widening of the cleared area to the northeast conceivably results from the sensitivity of the cap to very small energy inputs near its edge as defined by insolation, although other effects such as topographic slopes may be involved). The critical problem with this suggestion is the low probability of observing a ridge soon enough after its eruption that the heatflow is still enhanced. In our model, the peak heat flux is reached at the edges of the 34-km-wide zone defined above 7×10^4 years after emplacement of the dike. No impact craters superimposed on the ridges were detected by Voyager, but a crater-density age on the order of a few hundred million years has been estimated for the Tritonian maria [8,10].

If the tens of linear ridges observed were erupted over a similar time span, the probability of observing one within 10^5 years of its formation is only of the order of 1%. This difficulty might be overcome if openings in the permanent polar cap, once created, are able to persist because of the lower albedo of the exposed substrate. Additional, numerical modeling would in any case be of interest to determine how the result of our idealized model is modified by heat released from the surface flow, which is comparable in width to the zone defined above, and how sensitive the retreating polar-cap edge is to small energy inputs.

We have also considered the effects of warm-ice diapirism on the surface heat flux. We have *not* attempted to construct a consistent model of the chemistry of the diapirs and the buoyancy forces driving them; this is in fact problematic because of the similar densities of ammonia-water liquids and their coexisting solid phases [11]. Instead, we have modeled the ascent of diapirs with a "generic" buoyancy $\Delta\rho/\rho = 0.1$ and a temperature of 173 K, which is both the NH_3-H_2O eutectic temperature and the approximate temperature at the base of the lithosphere. Our results are very weakly dependent on the choice of these parameters. Because of the high ice viscosities and steep viscosity gradient in Triton's cold lithosphere, diapiric ascent proceeds not by Stokes-like flow, but by "thermal lubrication flow" [12], using its own heat to soften a thin layer of the ice around it. We have adapted the diapiric-ascent model of Kirk and Stevenson [3], incorporating this physics to Tritonian conditions. The model does not account for cooling of the diapir, so we stop the ascent after one thermal-diffusion time based on diapir radius. We find that the diapir typically ascends 1.0–1.5 of its radii before stopping (Fig. 2). Using a one-dimensional conduction model to calculate the enhanced heatflow above the center of the diapir, we find that a diapir with a radius of 70 km is required to double the ambient heat flux if $F_0 = 6.6\,\mathrm{mW\,m^{-2}}$, while a 100-km-radius diapir is needed to double our smaller heat flux. The corresponding thermal-diffusion times are 50 and 100 million years. Observation of the thermally active phase is thus much more probable than was the case for the linear ridges. We suggest that the three diffuse, roughly circular, low-albedo features at latitude 5° S, longitudes 25°–50° on Triton [8] may be the result of modification of the surface frost by diapiric activity. The features are roughly 50 km in radius, and at least two of them are clearly associated with mare-type cryovolcanism.

References Cited [1] Brown, R. H., *et al.*, submitted to *Science*. [2] Tyler, G. L., *et al.* (1989) *Science*, **246**, 1454–1458; Broadfoot, *et al.* (1989) *Science*, **246**, 1459–1465. [3] Kirk, R. L., and D. J. Stevenson (1987) *Icarus*, [4] Hobbs, P. V. (1974) *Ice Physics*, Clarendon Press, 357–361. [5] Scott, T. A. (1976) *Phys. Rep.*, **27**, 85–157. [6] Stansberry, J. A., *et al.* (1990) *Geophys. Res. Lett.*, **17**, 1773–1776. [7] Spencer, J. R. (1990) *Geophys. Res. Lett.*, **17**, 1769–1772. [8] Smith, *et al.* (1989), *Science*, **246**,, 1422–1449. [9] Kirk, R. L. (1990), *Lunar Planet. Sci.*, **XXI**, 631–632. [10] Strom, R. G., *et al.* (1990) *Science*, **250**, 437–439. [11] Croft, S. K., *et al.* (1988) *Icarus*, **73**, 279–293. [12] Morris, S. (1982) *J. Fluid Mech.*, **124**, 1–26.



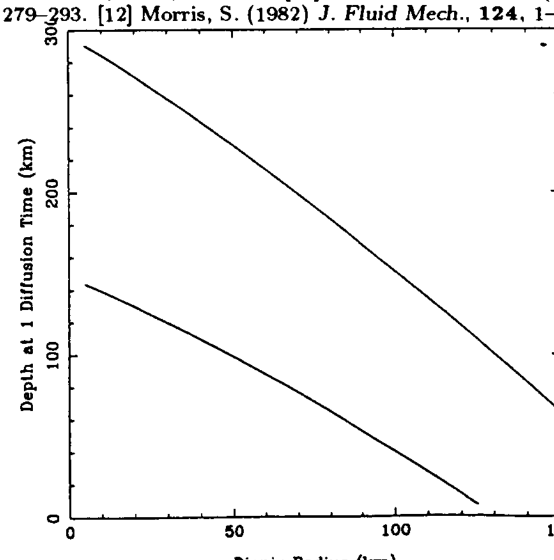


Figure Captions **Fig. 1** **Redistribution of internal heat flux as a function of cap thickness at pole. Solid curves: flux decrease at pole (relative to mean flux); dashed curves: twice flux increase at equator. Upper and lower curve sets correspond to maximum and minimum internal fluxes considered.** **Fig. 2** **Minimum depth of ascent of 173 K Tritonian diapir as a function of radius. Upper and lower curves correspond to minimum and maximum internal fluxes considered.**