Heat flow and depth to a possible internal ocean on Triton

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Abstract

The Raz Fossae, a pair of ≈ 15-km wide trough en echelon interpreted as grabens, can be used to propose an estimation of the depth to the brittle–ductile transition on Triton. This estimation may in turn give an idea of the thermal state of Triton's icy lithosphere when these features formed. Given the young age of its surface, the conclusions obtained could be roughly applicable to the present state of this satellite of Neptune. Considering water or ammonia dihydrate as possible components of the lithosphere and a feasible range of strain rates, it was estimated that surface heat flow is greater than that inferred from radiogenic heating, especially for a lithosphere dominated by water. Also, an internal ocean could lie at a depth of only ∼ 20 km beneath the surface. The presence over the surface of an insulating layer of ice of low thermal conductivity (e.g., nitrogen) or of regolith would only substantially alter these estimates if the effective surface temperature were considerably higher than the observed value of 38 K.

Keywords: Triton; Thermal histories

1. Introduction

The Raz Fossae, two prominent, ≈ 15 km wide, troughs en echelon is among the most outstanding tectonic features of Triton (Fig. 1), the large icy satellite of Neptune. These structures have been interpreted as grabens (Croft et al., 1995). Through simple geometric concepts, the maximum depth of faulting can be estimated from the width of a graben: it is generally assumed that the depth of intersection of the walls of a graben coincides with a mechanical discontinuity (e.g., Golombek, 1979; McGill and Stromquist, 1979). This discontinuity could correspond to a change in material composition or to the brittle–ductile transition depth. The latter marks the depth at which ductile (and temperature-dependent) creep starts to prevail over brittle failure as the dominant deformation mechanism. In this sense, it has been suggested that the width of Raz Fossae would indicate that the depth of faulting is related to the depth of the base of Triton's brittle lithosphere (Croft et al., 1995) in the time when this feature formed. In this paper, the width of the Raz Fossae is assumed to be related to the depth to the brittle–ductile transition, and so is used to calculate the local thermal state of Triton's icy lithosphere when these structures were formed by basically following the method described in Ruiz and Tejero (2000).

It is generally accepted that Triton's icy layer is mainly composed of water ice (Smith et al., 1989; McKinnon et al., 1995), though it has been argued using cosmochemical concepts, that ammonia–water system ices (e.g., ammonium dihydrate) might be a closer approximation for an icy satellite (e.g., Hogenboom et al., 1997). Hence, calculations are conducted both for water and ammonia dihydrate ices. The results obtained are used to estimate the depth to the top of a putative internal ocean at the time when the Raz Fossae formed. The possibility of an elevated effective surface temperature because of ices of low thermal conductivity, solid-state greenhouse or a regolith layer on the surface of Triton is also taken into account.

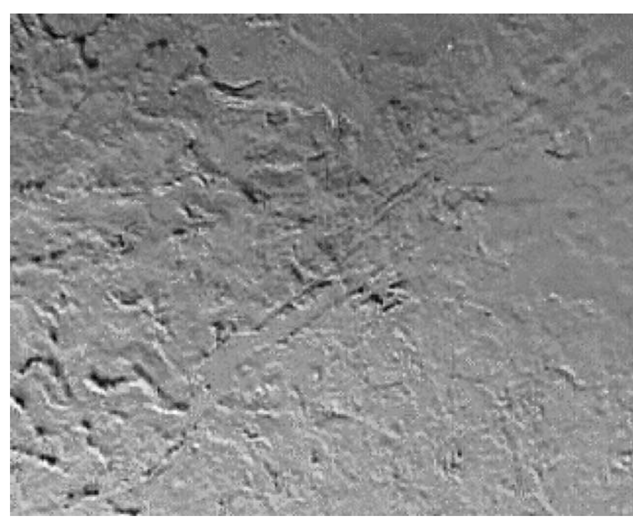


Fig. 1. The Raz Fossae, Triton, seen in a mosaic of images taken by the Voyager 2 spacecraft. These features are ≈ 15 km wide and are centered on 8° N, 21.5° E.

From the low density of impact craters, it may be inferred that the surface of Triton is relatively young (Smith et al., 1989). Recent age estimates give ⩽ 300–350 Myr (Stern and McKinnon, 2000; Zahnle et al., 2003). Such recent surface ages indicate that the results obtained herein could be roughly applicable to the current state of this notorious satellite.

2. Lithospheric heat flow when the Raz Fossae were formed

If the Raz Fossae are interpreted as grabens, then their maximum possible depth z is given by $(w/2)\tan\beta$ (e.g., Golombek, 1979; McGill and Stromquist, 1979), where w is the width of the graben and β is the fault's dip angle. Given the lack of experimental data, here I assume that the frictional properties of water ice (Beeman et al., 1988) are also valid (roughly at least) for ammonia dihydrate ice. The most favorable angle for normal faulting in water ice is $\approx 60°$ (Beeman et al., 1988). Thus, taking $w = 15$ km and $\beta = 60°$, gives $z = 13$ km. Terrestrial normal faults have typically dip angles of 55°–65° (Turcotte and Schubert, 2002); taking in account this range for the Raz Fossae gives $z \approx 11$–16 km.

In a cohesive material such as water ice, brittle strength at the base of a normal fault can be described as (Ruiz and Tejero, 2000)

$$(\sigma_1 - \sigma_3)_b = \frac{2B(\mu\rho g z + S)}{2\mu B + 1}, \tag{1}$$

where μ is the friction coefficient, $B = (\mu^2 + 1)^{1/2} + \mu$, ρ is the density, g is the gravity (taken as the surface value, 0.78 m s^{-2}, for Triton), and S is the material's cohesion. For water ice (whose frictional characteristics are assumed valid for Triton's entire icy lithosphere), $S = 1$ MPa and $\mu = 0.55$ (Beeman et al., 1988), and thus $B = 1.69$. The result does not vary much if the density is taken as that of very cold water ice (940 kg m^{-3}) or that of ammonia dihydrate (965 kg m^{-3}).

In turn, the ductile strength of ice is given by

$$(\sigma_1 - \sigma_3)_d = \left(\frac{\dot{\varepsilon} d^p}{A}\right)^{1/n} \exp\left(\frac{Q}{nRT}\right), \tag{2}$$

where A, p, and n are laboratory-determined constants, d is the grain size, $\dot{\varepsilon}$ is the strain rate, Q is the activation energy of creep, $R = 8.3145$ J mol^{-1} K^{-1} is the gas constant, and T is the absolute temperature. In planetary conditions, water ice creep can occur by superplastic flow, which is grain size-sensitive, or dislocation creep, which is independent of grain size ($p = 0$). The higher the temperature, grain size and differential strain, the greater is the relative contribution of dislocation creep. Since the grain size of Triton's lithosphere is unknown, I present calculations for both these water ice deformation mechanisms. For superplastic flow $Q = 49$ kJ mol^{-1}, $A = 0.0039$ MPa^{-n} m^p s^{-1}, $p = 1.4$, and $n = 1.8$ (Goldsby and Kohlstedt, 2001), and it is assumed $d = 0.1$–1 mm (a typical range used for icy satellites; e.g., Ruiz and Tejero, 2003). For dislocation creep $Q = 61$ kJ mol^{-1}, $A = 10^{5.1}$ MPa^{-n} s^{-1} and $n = 4$ (Durham et al., 1997). I also performed calculations for ammonia dihydrate, for which there is available the dislocation creep flow law: $Q = 107.5$ kJ mol^{-1}, $A = 10^{21.55}$ MPa^{-n} s^{-1}, and $n = 5.81$ (Durham et al., 1993).

The strain rate, which is unknown, is needed to resolve Eq. (2). I then made calculations corresponding to a feasible range of strain rates for planetary conditions 10^{-10}–10^{-20} s^{-1}. The upper limit, 10^{-10} s^{-1}, is approximately the strain rate tidally induced in the ice shell of Europa (Ojakangas and Stevenson, 1989). The lower limit, 10^{-20} s^{-1}, is an order of magnitude slower than strain rates attributable to thermal contraction on terrestrial planets (e.g., Anderson and Grimm, 1998). The mid interval value of 10^{-15} s^{-1} represents a strain rate that is considered typical in works about planetary bodies, including icy satellites (e.g., Golombek and Banerdt, 1990).

Equating Eqs. (1) and (2) gives the temperature at the level of brittle–ductile transition, T_{BDT}, which in turn allows calculation of surface heat flow dissipated by the satellite. The thermal conductivity of water ice is dependent on temperature according to $k = k_0/T$; in this case the heat flow is given by

$$F = \frac{k_0}{z} \ln\left(\frac{T_{\mathrm{BDT}}}{T_{\mathrm{s}}}\right), \tag{3}$$

where $k_0 = 567$ W m^{-1} (Klinger, 1980), and T_{s} is the surface temperature. There are some available measures of the thermal conductivity of ammonia-rich ice (Ross and Kargel, 1998; Lorenz and Shandera, 2001). It can be deduced from these measurements that the thermal conductivity of ammonia dihydrate is ~ 1 W m^{-1} K^{-1}, but a clear temperature dependence cannot be inferred; here I therefore use a constant value of 1 W m^{-1} K^{-1} for ammonia dihydrate, and in this case the heat flow is given by

$$F = \frac{k_{\mathrm{a}}(T_{\mathrm{BDT}} - T_{\mathrm{s}})}{z}, \tag{4}$$

where k_{a} is the thermal conductivity of ammonia dihydrate.

Table 1 summarizes the surface heat flow values obtained for $T_{\mathrm{s}} = 38$ K (Tryka et al., 1993) and both H_2O and $NH_3{\cdot}2H_2O$ lithospheres. Figure 2 shows the results for the nominal $\beta = 60°$ case. The surface heat flow due to radiogenic heating has been estimated at ~ 3.3 mW m^{-2} (Brown et al., 1991), assuming a chondritic composition for Triton's rock fraction. Table 1 and Fig. 2 indicate that heat flows obtained in this work for a H_2O lithosphere are much higher than for an $NH_3{\cdot}2H_2O$ one, and much higher than current heat flow due to radiogenic disintegration in the rock/metal core. In contrast, for a $NH_3{\cdot}2H_2O$ lithosphere the surface heat flow is lower, but still clearly higher than that expected due to radiogenic heating.

Table 1
Results for $T_{\mathrm{s}} = 38$ K and $\beta = 60°$ (55°–65°)

	10^{-10}–10^{-20} s^{-1}	10^{-15} s^{-1}
H_2O lithosphere		
Surface heat flow (mW m^{-2})	44–71 (35–86)	53–57 (43–67)
Putative ocean depth (km)	16–26 (13–32)	20–21 (17–26)
$NH_3{\cdot}2H_2O$ lithosphere		
Surface heat flow (mW m^{-2})	6.3–8.9 (5.1–10.8)	7.4 (6.0–9.0)
Putative ocean depth (km)	16–22 (13–27)	19 (15–23)

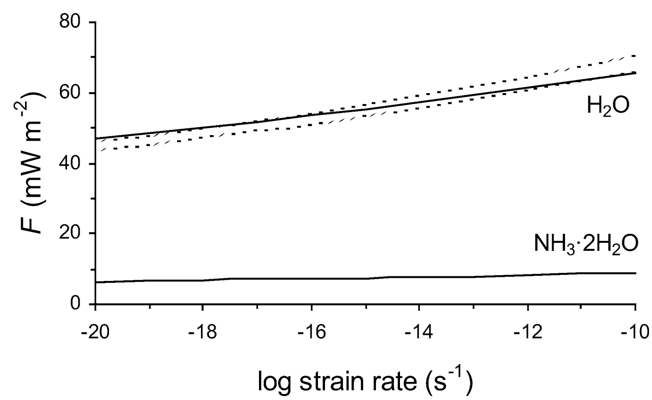


Fig. 2. Surface heat flow consistent with a brittle–ductile transition at the maximum depth of the Raz Fossae (≈ 13 km for a fault's dip angle of 60°), for H_2O and $NH_3{\cdot}2H_2O$ lithospheres, in terms of the strain rate. Solid lines represent ductile deformation due to dislocation creep. Dotted lines indicate ductile deformation due to superplastic flow in water ice of grain size 1 mm (upper curve) and 0.1 mm (lower curve).

3. Depth to a possible internal ocean

If Triton's ice mantle contains significant amounts of ammonia (a potent antifreezing), there could be an extensive liquid layer of ammonia–water inside this satellite (e.g., McKinnon et al., 1995). It is even possible there might be a liquid-water-dominated ocean within Triton. Indeed, it has been recently proposed (Ruiz, 2001) that, due to the non-Newtonian viscosity of water ice (the same is true for ammonia dihydrate ice), the onset of convection in the outer layers of icy satellites is much more difficult than previously thought, allowing outer shells stables against convection, and possible internal oceans to escape freezing. Tidal stresses can reduce the viscosity of ice (McKinnon, 1999), which would favor the onset of convection, but current tidal stresses are not important on Triton, and so this effect is not considered here.

In the absence of convection, the temperature profile of the outer ice layer is purely conductive. If one assumes an $NH_3 \cdot 2H_2O$ ice layer, then the depth to the ocean top would be given by

$$z_{\mathrm{ocean}} = \frac{k_{\mathrm{a}}(T_{\mathrm{m}} - T_{\mathrm{s}})}{F}, \tag{5}$$

where T_{m} is the melting temperature of $NH_3 \cdot 2H_2O$ ice, ~ 176 K (e.g., Kargel, 1992), a value which varies slightly with pressure (e.g., Hogenboom et al., 1997). In the case of a water ice layer (of variable thermal conductivity), the depth of the ocean top is given by

$$z_{\mathrm{ocean}} = \frac{k_0}{F} \ln\left(\frac{T_{\mathrm{m}}}{T_{\mathrm{s}}}\right). \tag{6}$$

Although T_{m} is pressure-dependent, given the low pressures of Triton's lithosphere, a constant value of 273 K was considered for T_{m}.

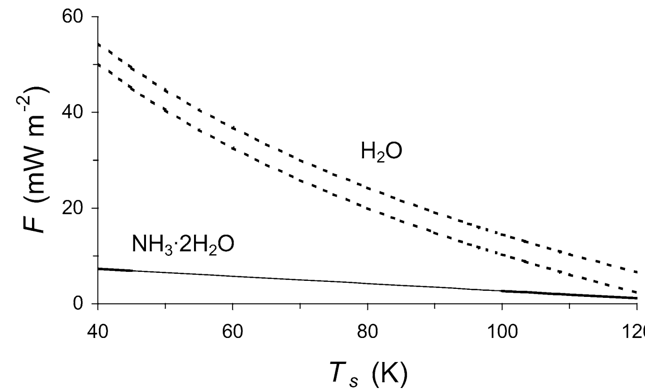
Table 1 shows the range of depths to the top of an internal ocean consistent with the heat flows obtained in Section 2 and the compositions considered. The depth to the top of the putative ocean is similar for both possible compositions of the ice layer. Thus, when the Raz Fossae formed, it is probable there was an ocean on Triton lying at a depth of ~ 15–30 km; for $\beta = 60°$ and a typical strain rate of 10^{-15} s^{-1}, the depth to the ocean would be ~ 20 km.

4. Effect of an insulating surface layer

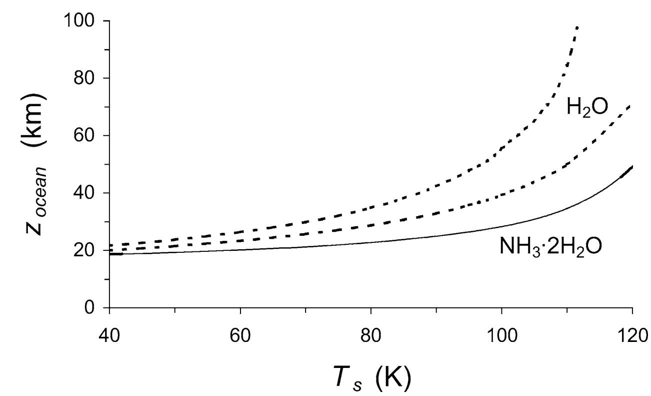
The presence on Triton's surface of ices of nitrogen, carbon monoxide, carbon dioxide, methane and water has been reported (e.g., Brown et al., 1995; Quirico et al., 1999), and their spatial distribution can be somewhat irregular. Ices other than water ice show low thermal conductivity (Ross and Kargel, 1998). For instance, the thermal conductivity of nitrogen (the dominant surface species) is 0.2 $\mathrm{W\,m^{-1}\,K^{-1}}$, or two orders of magnitude less than the conductivity of water ice at temperatures typical of Triton's surface. Consequently, a surface layer rich in low thermal conductivity ices would raise the temperature of the most superficial layers (i.e., the effective surface temperature) and contribute to a hotter interior (McKinnon et al., 1995). Similarly, such as it has been suggested for other satellites of the outer Solar System, the presence of an insulating regolith layer (e.g., Shoemaker et al., 1982) or a solid-state greenhouse effect (Brown and Matson, 1987; Matson and Brown, 1989) could considerably increase the effective surface temperature.

Thus, Fig. 3a shows surface heat flow deduced from the depth of Raz Fossae as a function of effective surface temperature. Calculations were made for $\beta = 60°$ and $\dot{\varepsilon} = 10^{-15}$ s^{-1}. As expected, heat flow diminishes with T_{s}, since the difference between T_{s} and T_{BDT} becomes smaller. For surface heat flow to be consistent with the predicted value of radiogenic heating, the effective surface temperature needs to be increased to at least ~ 90 K (for $NH_3 \cdot 2H_2O$) or to ~ 120–130 K (for H_2O).

Similarly, Fig. 3b shows the depth to an internal ocean as a function of effective surface temperature for the heat flows shown in Fig. 3a. The increase in T_{s} raises the interior temperature, but this effect is offset by the reduction in the value of F necessary to put the brittle–ductile transition to the base of the Raz Fossae. Hence, for a conductive ice shell, the depth to a possible ocean increases with T_{s}. If we consider a lower limit of 3.3 $\mathrm{mW\,m^{-2}}$, then the depth to an internal ocean should be $\leqslant 30$ km for an $NH_3 \cdot 2H_2O$ ice layer, and $\leqslant 130$–145 km for a layer of H_2O ice. Given the overall thickness of Triton's ice layer is estimated at ~ 350–400 km (Smith et al., 1989; McKinnon et al., 1995), the existence of an internal ocean seems inevitable in the absence of convection in the outer ice layer. Moreover, if the effective surface temperature is not increased by the insulating layer to the extent of several dozen degrees, the ocean would lie at a depth of ~ 20–30 km beneath the surface.



(a)



(b)

Fig. 3. (a) Surface heat flow necessary for put the brittle–ductile transition at the base of Raz Fossae, as a function of the effective surface temperature for a strain rate of 10^{-15} s^{-1}. The dotted lines indicate upper and lower limits for an H_2O lithosphere and the solid line a $NH_3 \cdot 2H_2O$ lithosphere. (b) Depth to a possible internal ocean underlying a conductive outer ice layer as a function of the effective surface temperature and heat flows shown in Fig. 2a. The increase in T_{s} raises the interior temperature but this effect is offset by the diminished value of F. The dotted lines indicate upper and lower limits for an H_2O lithosphere and the solid line a $NH_3 \cdot 2H_2O$ lithosphere.

5. Discussion and conclusions

The young surface of Triton's seems to indicate that this satellite is currently active (Stern and McKinnon 1999, 2000; Schenk and Sobieszczyk, 1999). Stern and McKinnon (2000) estimated that, even for their conservative age estimates, Triton has the highest resurfacing rates of the outer Solar System after Io and Europa, being of the same order as those calculated for Venus and the Earth's intraplate zones. These high rates of geological activity are consistent with the high heat flows obtained in Section 2. Such heat flows might be the remains of the intense heat generated by the capture of a heliocentric Triton by Neptune (see McKinnon et al., 1995, for a comprehensive review).

The young Triton's surface could also point to the existence of an internal ocean (Schenk and Sobieszczyk, 1999; Stern and McKinnon, 1999), which could in turn be a source of magmas for resurfacing. An ocean lying ~ 20 km below the surface, as estimated in Section 3, would be consistent with this rationale.

Further, the cantaloupe terrain of the tritonian Bubembe Regio may have formed via diapiric overturn of a layered crust ~ 20 km thick (Schenk and Jackson, 1993). Though this event precedes the opening of the Raz Fossae (Croft et al., 1995), this thickness is comparable to the depth to the internal ocean proposed here. Thus, the results of this works suggest that the cantaloupe terrain may have formed by diapiric overturn of the whole ice layer.

Finally, the results presented in Section 4 show that the presence on the surface of an insulating layer (low thermal conductivity ices or regolith) could only alter the overall conclusions of this work if it substantially increases the effective surface temperature. It is therefore feasible that Triton has retained part of the heat generated during its capture by Neptune and is able to sustain an internal ocean at a depth as low as 20 or 30 km below its surface.

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