THE GEOLOGY OF THE ROCKY BODIES INSIDE ENCELADUS, EUROPA, TITAN, AND GANYMEDE.

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Introduction: The icy satellites of Jupiter and Saturn have been the subjects of substantial geological study. Much of this work has focused on their outer shells [e.g., 1–3], because that is the part most readily amenable to analysis. Yet many of these satellites feature known or suspected subsurface oceans [e.g., 4–6], likely situated atop rocky interiors [e.g., 7], and several are of considerable astrobiological significance. For example, chemical reactions at the rock–water interface might support the development of chemoautotrophic habitats there [e.g., 8], and hydrothermal systems and even seafloor volcanism might be present within some moons [e.g., 9]. Of course, the silicate interiors of these bodies are not directly accessible, and so their geological properties remain largely unknown. We therefore combined rock mechanics techniques with remotely sensed geophysical data for several icy satellites, to place first-order estimates on the mechanical and geological properties of their rocky portions. Here, we report initial results for Enceladus, Europa, Titan, and Ganymede.

Methods: Our approach begins with calculating how acceleration due to gravity, g, and overburden pressure, p, changes as a function of body radius, r. The change in g with depth is given by

$$\delta g/\delta r = 4\pi G\rho_r - 2(g/r), \qquad (1)$$

and p varies with depth by

$$\delta p/\delta r = -\rho(g), \qquad (2)$$

where G is the gravitational constant and ρ is material density [10]. Next, the change in porosity within the rocky interiors of these moons as a function of g can be estimated from (1) with the relation

$$\rho_\infty/(1+\{V_0/(1-V_0)\}\exp\{-\lambda\rho_\infty gh\}), \qquad (3)$$

where ρ_∞ is the density of compacted rock, V_0 is the void space fraction at the surface of the rock, λ is a constant, and h is depth [11]. Moreover, fault strength can be characterized, under the assumption that p (from (2)) denotes the maximum principal compressive stress (i.e., σ_1), with the equations

$$\sigma_1/\sigma_3 = (S_v-P_p)/(S_h-P_p) \leq (\{\mu^2+1\}^{0.5}+\mu)^2 \qquad (4)$$

and

$$\sigma_1/\sigma_3 = (S_H-P_p)/(S_v-P_p) \leq (\{\mu^2+1\}^{0.5}+\mu)^2 \qquad (5)$$

for normal and thrust faults, respectively; here, σ_3 is the minimum principal compressive stress, S_v is the vertical stress, S_h and S_H are the minimum and maximum horizontal stresses, respectively, P_p is pore fluid pressure (found from (3)), and μ is the coefficient of friction [12]. Finally, because equations (4) and (5) assess failure in the brittle domain, we also considered ductile deformation with the relation

$$\dot{\varepsilon} = C_1\sigma^n\exp^{-E/RT}, \qquad (6)$$

where $\dot{\varepsilon}$ is strain rate, C_1 is a constant, σ is deviatoric stress, n is the stress exponent, E is activation energy, R is the universal gas constant, and T is temperature [13].

Enceladus: With published values for the interior structure of Enceladus [14], we find that the pressure at the ocean floor is about 7 MPa, comparable to that at the bottom of the North Sea on Earth. The very low gravitational acceleration at Enceladus results in a modest porosity (e.g., 10%) at the rock–water interface remaining above zero at the center of the rocky interior. Were this interior permeable, then it is likely that the entire rock volume of Enceladus is saturated with water—or, at least, has been entirely serpentinized by earlier hydrothermal activity [e.g., 8]. Such a finding of pore space through to the rocky center supports recent inferences that sustained geological activity inside the moon has been facilitated by a largely unconsolidated silicate interior [15]. We show estimates for the brittle strength of the interior in **Figure 1**.

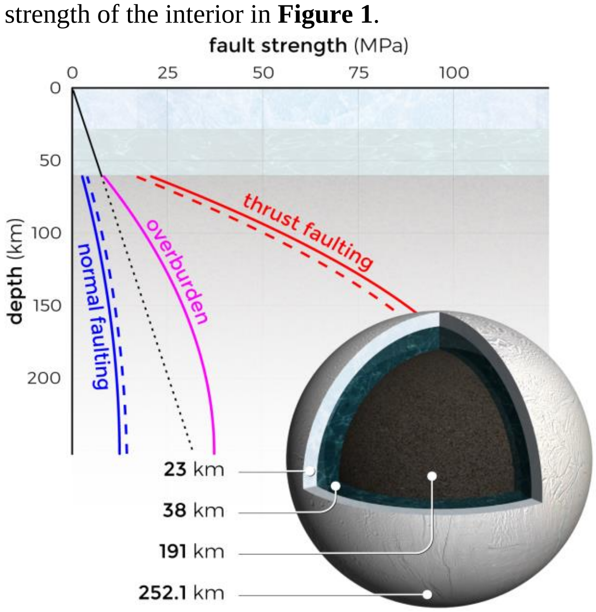

Figure 1. *Strength profiles for normal and thrust faults, and hydrostatic and lithostatic stress, for Enceladus' interior [14].*

Europa: The deeper ocean at Europa [7,16], coupled with that body's higher *g*, results in an ocean floor pressure of about 210 MPa, much greater than that at Enceladus (and equivalent to almost twice that at the bottom of the Marianas Trench). Further, even moderately high porosities (~40%) at the ocean floor reduce to zero within but a few kilometers; pore space, and thus hydrothermal alteration, likely does not prevail below this depth. As for Enceladus, we calculated likely fault strength profiles for the brittle rocky portion of Europa, but also derived a strength envelope for a putative, ductile lower part of the moon's rocky interior. (We assumed minimum and maximum values of $\dot{\varepsilon}$ of 10^{-20} and 10^{-14} s^{-1}, respectively, with *T* calculated from thermal gradients bracketed by 5 K/km and 20 K/km.) We find the brittle–ductile transition (BDT) within Europa—the maximum depth to which faults can penetrate before strain is accommodated via crystal plasticity and other ductile deformation mechanisms—to range from over 800 km (for the lowest thermal gradient and highest strain rate) to as little as 160 km (for the highest thermal gradient and lowest strain rate).

Titan and Ganymede: As dramatically larger bodies, the pressures at the Titanean and Ganymedean rock–water/ice interfaces are substantially higher than those at Enceladus and Europa. For the likely interior structures of Titan [17] and Ganymede [18], this pressure is about 1.1 GPa and 1.2 GPa, respectively. Any porosity that might exist at the rock surface is therefore negligible at a depth of only a few hundred meters. Depths to the BDT within Titan and Ganymede (calculated for the same values as for Europa) are very similar, ranging from about 780 km (lowest thermal gradient and highest strain rate) to 150 km (highest thermal gradient and lowest strain rate). These results suggest that the mechanical stratigraphy of the rocky bodies inside Europa, Titan, and Ganymede may be broadly the same.

Outlook: The results we report here place broad constraints on the mechanical behavior and properties of the silicate interiors of several notable icy satellites. With our calculated strength profiles, for example, it becomes possible to determine whether faulting in response to diurnal or non-synchronous rotation tidal stresses [e.g., 2] has affected these interiors. Other sources of stress include global contraction, a process driven by interior secular cooling [19] that has, for instance, profoundly influenced the geological history of Mercury [e.g., 20], and has likely operated on Mars [21] and the Moon [19,22]. Of note, the rocky portions of Titan and Ganymede are, *by themselves*, about the same size as the Moon [17,18], a world that has experienced a sustained history of volcanic and tectonic activity. It is therefore plausible that similar histories have been recorded on the seafloors of Titan and Ganymede, with fault-induced topography, volcanic and hyaloclastite deposits, and even considerable sedimentary accumulations (albeit without a terrigenous and/or biogenic component) present there. Moreover, the ability of faults to penetrate to substantial depths within these bodies could enhance the circulation of seawater through, and the hydrothermal alteration of, the upper crust as well as facilitate the ready ascent of magma.

This approach can be applied to other rocky bodies ensconced in water/ice, whether fully differentiated or not, such as Dione, Tethys, Pluto, Charon, Triton, or any of the large moons of Uranus. Additionally, future measurements of these worlds by visiting spacecraft will refine our input parameters, such as those planned for the Europa Clipper mission [23]; those missions could even test for evidence of geological activity at the ocean floor [e.g., 24]. More broadly, the likely complex geological properties and histories of these rocky interiors warrants them being considered planetary bodies in their own right. From this perspective, the Solar System contains not a handful of terrestrial planets, but almost two dozen such worlds.

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