RISE OF DEEP MELT INTO GANYMEDE'S OCEAN AND IMPLICATIONS FOR ASTROBIOLOGY.

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Introduction: Results from the Galileo magnetometer point toward the presence of a layer of liquid water within Ganymede, presumably near the 170 km depth of the minimum ice melting temperature [1]. Unlike an ocean within Europa, which may be in direct contact with hydrothermal systems of a rocky sea floor, Ganymede's ocean is sandwiched between layers of ice I and ice III, so it is seemingly isolated from the chemical nutrients that might sustain a biota. However, it is plausible that large magmatic events at Ganymede's deep rock/ice interface could generate pockets of water melt, which may then rise buoyantly through high-density ice phases of the ice mantle, carrying nutrients to Ganymede's ocean. Here we introduce an investigation of the conditions under which such melt pockets could reach the ocean.

Volcanism and Rising Water: Gravity data suggests that Ganymede (radius 2631 km) is differentiated into an ice mantle approximately 900 km thick, a rocky outer core ~400 to 1300 km thick, and iron inner core with a radius of 400 to 1300 km [2]. From the surface downward through the ice mantle, Ganymede is inferred to consist of concentric shells of ice I, briny liquid water, and high pressure polymorphs ice III, V, and VI. Water is buoyant relative to all of these high-density ice polymorphs, which have mean densities of 1.14, 1.23, and 1.31 g/cm^3, respectively [3].

An individual magmatic event at the boundary between Ganymede's rocky interior and the overlying ice is assumed to melt out a single dome of water. Each water dome will rise as a Rayleigh-Taylor instability through Ganymede's icy mantle, following the path of the ice solidus through pressure-temperature space. Although it is plausible that many small instabilities will form, the rise time of a small plume is much larger than its thermal diffusion time, so a small instability will stall in the mantle and not reach the ocean, while a large one may reach the ocean.

As the instability moves up through the ice mantle, its size is controlled by how quickly it moves and loses heat to its surroundings. To solve for the radius of the instability as a function of time, we begin by setting the rate of change of the heat of the instability equal to the temperature gradient divided by the characteristic distance over which the heat diffuses:

$$(-4\pi R^2 \rho L_{ice})\frac{dR}{dt} = \frac{4\pi R^2 k\Delta T}{\left(\frac{\kappa R}{v}\right)^{(1/2)}}$$

where R is the radius of the instability, ρ is the density of ice, L_{ice} is the latent heat of fusion of ice, k is the thermal conductivity of ice, κ is the thermal diffusivity, and v is the Stokes velocity, given by:

$$v = \frac{2g\Delta\rho R^2}{9\eta}$$

Here, Δρ is the density difference between the instability and the surrounding ice, and η is the viscosity of ice. The acceleration of gravity, g, is evaluated at an average radius of the mantle, approximately 600 kilometers in depth.

Rheological data on the high pressure polymorphs of ice suggest that ice V, which occupies a ~100 kilometer layer beginning at a depth of approximately 400 km below the surface between layers of ice III and VI, is viscous relative to other polymorphs at similar temperatures [3]. Melting point viscosities for ice V are typically 10^{15} to 10^{17} Pa•s at a stress of 0.3 MPa [3]. This is two orders of magnitude higher than melting point viscosities of ice VI at the pressures and temperatures relevant to the interior of Ganymede, and 3 orders of magnitude greater than the nominal ice III viscosity [3]. Hence, the ice V layer is the most difficult layer for the instability to travel through. To be conservative in our criterion for success, we adopt 10^{16} Pa•s as an average melting point viscosity for all of Ganymede's ice mantle. The effects of the viscosity changes at the phase boundary layers will be examined in future efforts.

As a whole, Ganymede's ice mantle at present has Rayleigh number exceeding the critical Rayleigh number for the onset of solid state convection [4]. Our calculation assumes that the ice mantle has a steady state temperature of approximately 233 K and does not take into account the effects of the phase changes from ice V to ice III or ice V to ice VI. Phase changes with large Clapeyron slopes (dP/dT) can either enhance or retard convection and rising plumes [4]. Since the ice V-III and V-VI phase changes have dP/dT of approximately zero, this effect is small and can be neglected.

The temperature of a water instability will follow the solidus as the instability rises, and to simplify calculations, we take an average value for the temperature difference ΔT between the instability and the ambient ice. Both the buoyancy of the plume and the rate at which it conducts its heat away are dependent upon the temperature difference. As we will see, the plausible range in ΔT does not change the minimum eruption size by more than an order of magnitude.

To be successful, the instability must rise approximately 750 kilometers to travel from Ganymede's rocky core to the liquid water layer.

The equation for the radius of the instability as a function of time is given by:

$$R = \left(2R_o^{(1/2)} - t\left(\frac{k\Delta T}{\rho L}\right)\left(\frac{2g\Delta\rho}{9\eta\kappa}\right)^{(1/2)}\right)^2$$

The height that the plume reaches can be found by integrating the Stokes velocity with respect to time, the upper limit of integration being set by the time at which the radius of the plume goes to zero. In order to estimate the size of eruption that could produce a successful plume, we seek solutions where the final height exceeds 750 kilometers, given different values of R_o.

To calculate the magmatic mass m from the value of R_o, we assume that some percentage, f, of the heat released from the hot magma is used to melt out ice:

$$f[(mC\Delta T_m) + (m \times L_m)] = \frac{4}{3}\pi R_o^3 \rho \times (L_{ice})$$

Here, L_m is the latent heat of fusion of magma, and L_{ice} is the latent heat of fusion of ice, and ΔT_m is the temperature difference between the magma and ice. The contributions from the first and second terms on the left hand side are the same, to within an order of magnitude, since the heat capacity of a typical magma is 10^2 J/kg•K, ΔT_m is 10^3 K, and the latent heat of fusion of the magma is estimated to be $3x10^5$ J/kg. Since the heat of fusion of ice is approximately $3x10^5$ J/kg, the mass of the eruption is roughly equal to the mass of ice melted, divided by f. For calculations here, the efficiency factor f was taken to be 0.25.

Conclusions and Results: The results of the calculations are summarized in Figure 1. In our calculations, successful instabilities have a minimum initial radius R_o ~ 0.6 km, and correspond to magma eruption volumes on the order of 1 km^3. The amount of heat contained within a 1 km^3 eruption is approximately 10^{18} Joules. Varying the temperature of the ice mantle changes the required minimum eruption size from 0.5 km^3 to 5 km^3 for ΔT from 5 to 50 K, respectively.

Volcanic eruptions on Io similar to eruptive fissures in Hawaii have been estimated to erupt material at rates of up to 7 m^3 s^{-1} m^{-1} [5]. For Ganymede, given a reasonable fissure length of meters to 10 km, this corresponds to an eruption rate of a few 10's to 10^4 m^3/s, implying an eruption time of 3 hours to 16 days to make a plume of water of sufficient size to reach the ocean.

These calculations do not yet take into account potentially important effects of salts or other contaminants in the ice and water, and do not consider the detailed thermal structure of Ganymede's ice mantle, notably the possibility of multi-layered convection. Nevertheless, they show the feasibility that nutrient-rich water can be transported to Ganymede's thin ocean, opening what had been previously thought an isolated system, and perhaps helping to sustain a primitive biosphere. Further calculations are necessary to fully characterize the behavior of water plumes as they rise through Ganymede's ice mantle.

References: [1] M. G. Kivelson, et. al. (2000), The permanent and inductive magnetic moments of Ganymede, *Icarus*, submitted. [2] J. D. Anderson, et. al. (1996), *Nature*, 384, 541-543. [3] W. Durham, et. al. (1997), *JGR*, 102, 293-302. [4] W. B. McKinnon. (1998), in *Solar System Ices*, B. Schmitt et al., eds. [5] J. W. Head and L. Wilson (2000), Lava fountains from the 1999 Tvashtar Catena fissure eruption on Io: Implications for dike emplacement mechanisms, eruption rates and crustal structure, *JGR* submitted.

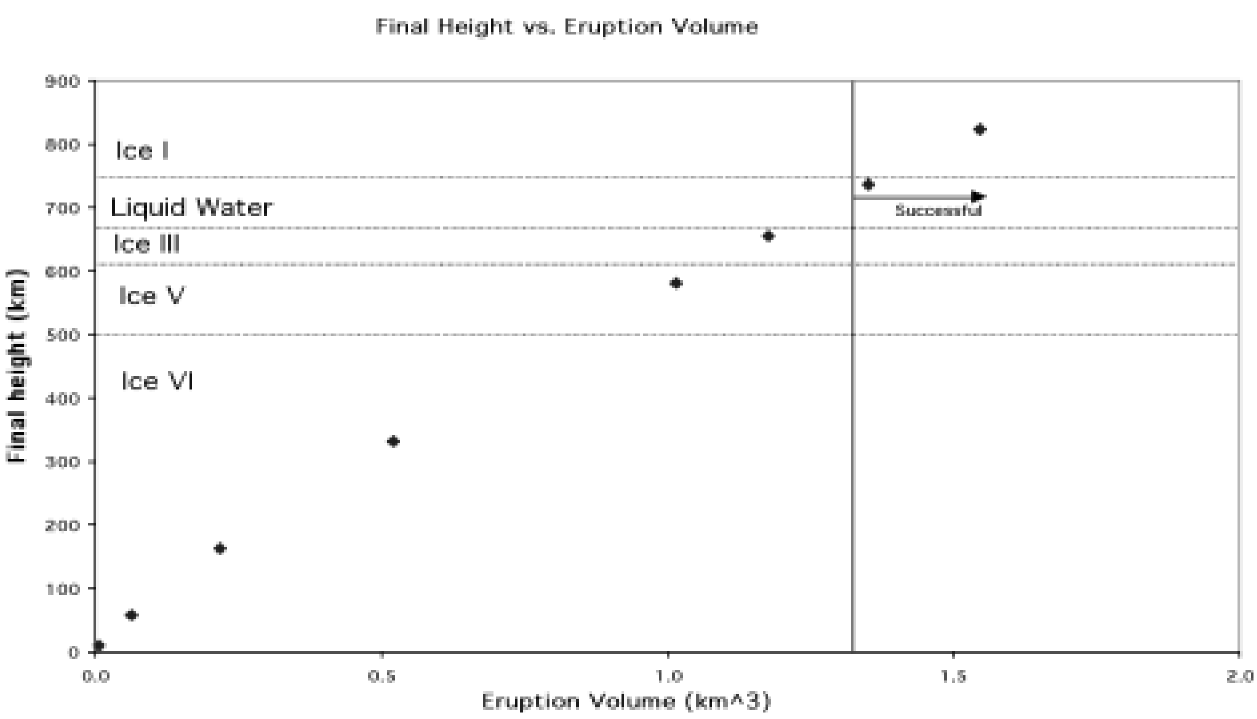

Figure 1. Final rise height for water plumes created at the rock/ice interface within Ganymede, as a function of the volume of a silicate magmatic erupted at this interface, hypothesized to create the water melt. Silicate eruptions of volume >1.4 km^3 can produce water pockets with initial radii >0.6 km, large enough to rise successfully (i.e., without entirely freezing) from the ice/rock interface (height = 0), through the intervening high-density ice phases, and into Ganymede's ocean. Such water plumes could conceivably supply nutrients to a primitive biosphere. Phase change locations are approximate.