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Pillow lava on Titan: expectations and constraints on cryovolcanic processes

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Abstract. Titan offers a fascinating environment in which to consider cryovolcanic processes: the Cassini/Huygens mission offers prospects for revealing cryovolcanic landforms, which may differ substantially from those on the other icy satellites owing to the effect of Titan's dense atmosphere. Various aspects of possible cryovolcanism on Titan are investigated in an attempt to predict what might be discovered. The thermal and stress environment on Titan is considered, and likely eruption rates and styles investigated; it is found that volcanic landforms are likely to be small, and "ash" cones are unlikely under present atmospheric conditions on Titan. Cooling rates for likely cryomagmas are investigated, and the possibility of rapidly-quenched pillow lavas is pointed out. Heat flow considerations limit the present resurfacing rate to $< 2 \times 10^{-3}$ m yr^{-1}: if atmospheric methane is buffered against photolysis by volcanic resupply, then the resurfacing rate is $> 1.5 \times 10^{-5}$ m yr^{-1}. Copyright © 1996 Elsevier Science Ltd

Introduction

Volcanism on Earth is usually based on silicate lavas. Water is not usually thought of as a volcanic fluid, although some volcano-like features can be found where water is vented into cold environments (e.g. the Lake Superior "icefoots"—Marsh *et al.* (1973)). Water-based volcanism on the icy satellites (hence "cryovolcanism") began to receive significant attention during and following the Voyager missions, which discovered many cryovolcanic features. Theoretical work following that mission's discoveries has focused on a number of particular features, including Ganymede's grooved terrain (Allison and Clifford, 1987) and topographic domes (Squyres, 1980), the resurfacing of Europa (Crawford and Stevenson, 1988; Squyres *et al.*, 1983) and flow features on Ariel and Miranda (Jankowski and Squyres, 1988; Schenk, 1991). The Galileo and Cassini missions promise to offer close looks at a number of icy satellite surfaces, giving us an opportunity to study cryovolcanic features more closely.

No such features have yet been seen on Titan: its surface was obscured from Voyager by the optically thick haze in the atmosphere. Since methane in Titan's atmosphere would be depleted in $\sim 1\%$ of the age of the solar system, it has been natural to speculate that methane might be released continuously or episodically from the interior by volcanism. As noted in Lorenz (1993), cryovolcanism on Titan may be very different from that on other satellites, as the presence of the thick (1.6 bar) atmosphere will affect the vesiculation of bubbles in a cryomagma, the distribution of tephra, and the cooling of lavas. Here I examine these questions in detail, in order to make predictions regarding the nature of cryovolcanism on Titan, if it indeed occurs.

The model of cryomagma transport from the interior to the surface is that of a fluid-filled crack, propagating upwards and sealing behind itself—see Stevenson (1982a,b) and Wadge (1980) and references therein.

Thermal environment

Volcanic activity transports heat from the interior of a body to its exterior. Clearly, the amount of heat that must be transported is instrumental in considering how much volcanic activity may occur. Titan is sufficiently large that during accretion, much of it may have melted. The interior (Stevenson, 1992; Lunine and Stevenson, 1987) may still contain a substantial layer of water–ammonia liquid (perhaps 300–400 km thick, Cynn *et al.* (1988)). This material represents a prime candidate for a cryomagma (see, e.g. Lewis, 1971; Kargel, 1990).

Titan's radiogenic heat production may be estimated as $\sim 4 \times 10^{11}$ W (Schubert *et al.*, 1986). Residual loss of

accretional heating (and the energy of differentiation) may provide an additional (but probably small at the present epoch) component.

In addition, the non-zero (0.029) eccentricity of Titan's orbit leads to some internal tidal dissipation. However, even for a "volatile-rich" interior model, with a liquid water–ammonia mantle, tidal dissipation is only $\sim 5 \times 10^{10}$ W (Sohl et al., 1995). Thus the radiogenic heating is the dominant element of the present day heatflow.

No attempt will be made here to model the transport of heat in Titan's interior, since such modelling is fraught with many uncertainties in ice rheology and starting conditions. Further, coupled orbital/thermal evolution of satellites can lead to large variations in tidal heat flow (Ojakangas and Stevenson, 1986). In this paper, I consider only present-day steady-state conditions.

However, a useful guide to the possibility of present-day volcanism is the depth of the 176 K isotherm, where water–ammonia could be in the liquid state. Assuming a geothermal temperature gradient compatible with the heat flow above and the thermal conductivity of solid water ice, this depth is about 30 km (Sohl et al., 1995), although if there was an insulating regolith layer present, the 176 K isotherm could be shallower.

Physical properties of cryomagmas

In this paper I assume that water–ammonia fluid is the cryomagma at work on Titan. The role of ammonia and other volatiles in outer planets was noted by Lewis (1971, 1972) and others: see also Consolmagno and Lewis (1978).

Key properties of a magma are the density contrast between it and the "country rock" and its viscosity. The density contrast between pure water–ammonia liquids and ices is investigated in Croft et al. (1988)—they found that the liquid phase is only marginally buoyant ($\Delta\rho \sim 10$ kg m^{-3}) near the peritectic. In this paper I assume $\Delta\rho \sim 100$ kg m^{-3}, since the crustal material may be contaminated with silicate material from impactors, and there may be small bubbles in the magma (see later). In general, eruption rates, crack lengths, etc. are only modestly sensitive to this parameter anyway.

The viscosity of a magma or lava is a prime determinant of the surface morphology of volcanic features. The viscosity of water–ammonia liquids was investigated extensively by Kargel: the key reference is Kargel et al. (1991), although there is also much material in Kargel (1990). These references give viscosity $\mu \sim 1$–10 Pa s for crystal-free water–ammonia liquids above the peritectic, to of order 10^3–10^4 Pa s for supercooled liquids and slurries (up to 10^6 Pa s for heavily-laden slurries). Arakawa and Maeno (1994) recorded viscosities in the range 10^7–10^{10} Pa s for partially molten water–ammonia ice. Kargel et al. (1991) also investigated water–ammonia–methanol fluids, and recorded viscosities somewhat higher than those for pure water–ammonia.

The chemistry of water–ammonia reactions with other compounds has been discussed by Thompson and Sagan (1992), Kargel (1992) and Engel et al. (1994).

For the latent heat of fusion of water–ammonia, I use 100 J g^{-1}: Croft et al. (1988) give ~ 132 J g^{-1} for ammonia dihydrate, while Yarger et al. (1994) measure up to 80 J g^{-1} for the enthalpy of crystallization of 20% ammonia solution (the exact value is sensitive to the exact composition, and the presence of dissolved impurities). I take the specific heat capacity as 1000 J kg^{-1}K^{-1}.

Titan's environmental parameters that are relevant are the surface gravity (1.35 m s^{-2}), the surface temperature (94 K) and the surface atmospheric pressure (1.6 bar).

Stress environment and crack propagation

The viscosity of an ice crust (Stevenson, 1982b) is likely to be too high to allow diapiric ascent of magma to the surface, so the most plausible way of escape from the interior is by fluid-filled cracks.

The varying distance from Saturn to Titan leads to crustal stresses. The maximum stress $(\sigma_n - \sigma_t)_{max}$ may be estimated as (Squyres and Croft, 1986)

$$(\sigma_n - \sigma_t)_{max} = \frac{24E(\Delta r/r)}{7(5+v)} \qquad (1)$$

where E is the Young's modulus of the crust, r the planetary radius, Δr the difference between planetary radii (due to the tidal distortion) and v the Poisson ration (~ 0.3). For Titan, with $r \sim 2575$ km, the equilibrium tidal shape has differences in radii of up to 509 m (Zharkov et al., 1985). However, for crack formation in the geological present, we require that component of Δr which is changing: since the tidal stress is proportional to the third power of distance from Saturn, the effective Δr is 509 m times $3e$, where e is the orbital eccentricity. Thus, $\Delta r \sim 50$ m and taking $E \sim 5$ GPa, the stress is therefore ~ 25 kPa, considerably less than the tensile strength of ice (~ 1–10 MPa). Thus tidal stress alone cannot generate cracks in the crust.

The extensional stress regime in a crater floor, as it undergoes viscous relaxation might provide a favorable environment for crack formation at the surface. Other effects, such as the emplacement of a large load on the surface, might lead to extensional stress at the base of the lithosphere. Once a crack forms, the influence of a buoyant fluid in it, or the cyclic tidal stresses, may cause it to propagate: here I consider only buoyancy-driven cracks forming at the base of the lithosphere.

According to the theory of crack propagation of Stevenson (1982a), the ascent velocity v of a fluid-filled crack is

$$v \sim 0.1(g\Delta\rho w^2/12\mu)^{0.857}(\kappa Z/w^2)^{0.143} \qquad (2)$$

where g is local gravitational acceleration, $\Delta\rho$ the density contrast, w the crack width, μ the viscosity of the magma, κ the thermal diffusivity of the surrounding rock, and Z the vertical length of the crack. A plausible upper limit for w is Δr, although smaller cracks are presumably more probable.

For Titan parameters ($g = 1.35$ m s^{-1}, $\Delta\rho = 100$ kg m^{-3}) ascent velocities typically of the order of a few cm s^{-1}, but up to about 1 m s^{-1} for large cracks
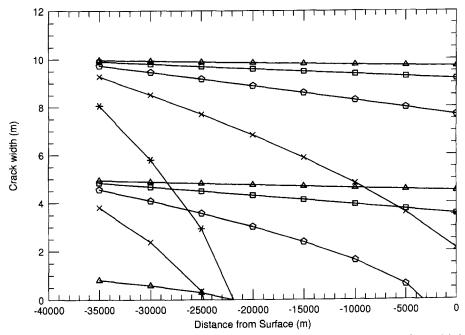
Fig. 1. Closure of cracks during ascent. Cracks are initiated at 40 km below the surface with initial widths of 10, 5 and 1 m and shrink by freezing as they ascend. Crack length Z is 5 km: magma viscosities are 10 (triangles), 100 (squares), 1000 (pentagons), 10^4 (crosses) and 10^5 Pa s (stars). Curves for 1 km long cracks are similar to those shown but for viscosities 1 decade higher (e.g. 1 km crack with 100 Pa s magma shrinks in width by the same amount as a 5 km crack with 1000 Pa s magma)

($w \sim 10$ m). Thus, from the initiation of a crack at the base of the lithosphere (~ 50 km down) to its arrival at the surface is $\sim 10^5$–10^6 s, or a day to a week or so.

It can be shown that, except for the thinnest, slowest cracks, the cryomagma is able to ascend without freezing. Crawford and Stevenson (1988) give the following expression for the rate of deposition of material on the face of an ascending crack

$$\sigma \Delta H \sim \rho C_p \Delta T (\kappa \tau)^{0.5} \qquad (3)$$

where σ is the mass per unit area of material frozen out, ΔH the latent heat of freezing, C_p the specific heat of the magma, ΔT the temperature difference (say ~ 50 K) between magma and bedrock, κ the thermal diffusivity (here $\sim 10^{-6}$ m^2 s^{-1}) and τ the residence time (which I estimate at $\sim 10^4$ s, the time for a large (~ 10 km tall, 10 m wide) crack to propagate its own vertical length (see above)). This expression yields about 50 kg m^{-3}, or an equivalent thickness of about 5 cm. Thus, for crossing a lithosphere of 3–10 crack lengths, only a few tens of centimeters of solid "rind" forms on the crack.

More formally, the freezing of a fluid-filled crack or dike may be considered using the transcendental equations given in Turcotte and Schubert (1986). These yield essentially (within $\sim 20\%$) the same result.

Figure 1 shows the evolution of crack width as a crack ascends—it is seen that from the base of the lithosphere (say ~ 40 km depth) cracks with lengths of 5 km and widths of 10 m can propagate to the surface for fluid viscosities of $< 10^3$ Pa s: 5 m wide cracks can do so if $\mu < 10^2$ Pa s. These propagations are for $\Delta \rho \sim 100$ kg m^{-3}: for $\Delta \rho \sim 10$ kg m^{-3} the corresponding viscosity limits are 10 times lower.

Table 1. Eruption rates for Titan cryomagma: units are m^2 s^{-1}: m^3 per second per unit length of crack

Crack width (m)	Eruption rate (Q/L) $\mu = 10$ Pa s	Eruption rate (Q/L) $\mu = 10^3$ Pa s
0.1	10^{-4}	2×10^{-6}
1.0	3×10^{-2}	6×10^{-4}
10	8.4	1.6×10^{-1}

Eruption rate

The eruption rate Q of a magma in m^3 s^{-1} from a crack of horizontal length L may be estimated by the formula (Wilson and Head, 1983a)

$$Q \sim w^2 L (wg\Delta \rho - 2\gamma)/12\mu. \qquad (4)$$

Here I ignore the yield strength γ of the bedrock (ice). Table 1 shows eruption rates per unit length of crack (Q/L) from this formula for Titan and typical water–ammonia rheologies of 10 and 1000 Pa s. Eruption rates for more viscous magmas would be smaller, but in any case, cracks filled with more viscous magmas are less likely to make it to the surface before freezing.

Eruption style

The eruption style of a volcano depends on the eruption rate, the viscosity and volatile content of the magma.

Magma disruption occurs, leading to "cyroclastic" eruptions (where molten or partially molten fragments of

ice are ejected from the vent), if the gas content of the magma exceeds 75% by volume (Wilson and Head, 1983a), or

$$3(1-n_e)/\rho_1 = n_e R_0 T/mP \qquad (5)$$

with n_e the mass fraction of the volatile, ρ_1 the magma density ($\sim 1000 \, \text{kg m}^{-3}$), R_0 the universal gas constant ($8.314 \, \text{J mol}^{-1}$), m the relative molecular mass of the volatile (here assumed methane, $0.016 \, \text{kg mol}^{-1}$), T the magma temperature (176 K) and P the surface pressure ($1.6 \times 10^5 \, \text{Pa}$), yields $n_e > 5 \times 10^{-3}$ for magma disruption.

However, the methane content of a water–ammonia magma is likely to be very low as methane is soluble to only about 5×10^{-3} mole fraction in water at 300 bar pressure (Lunine and Stevenson, 1985), corresponding to ~ 30 km depth on Titan (similar figures for the solubility of methane in water and in liquid ammonia may be found in Figs 22 and 24 of Kargel (1990)). Thus a magma column is barely likely to disrupt in Titan's present atmosphere.

Kargel and Strom (1990) showed that on Triton a cryo-magma would explosively disrupt at depths of a few tens or hundreds of meters, forming a spray of cryoclastic tephra and a maar-like explosion crater. However, this occurs due to the essentially unconfined expansion of the cryomagma volatile component (methane) into Triton's minimal atmosphere: on Titan, the expansion is limited by the significant atmospheric pressure. Thus, unless the volatile content of the magma can be significantly increased beyond 5×10^{-3}, explosive eruptions and ash clouds cannot form on Titan under present atmospheric conditions.

One possible way of adding "fizz" to a magma is if it ascends through a crack that was previously ascended by liquid methane which formed a layer of clathrate. (Note that the initial bulk composition of the crust cannot be clathrate as (Lunine and Stevenson, 1987) it had melted in the late stages of accretion: clathrate has to have been formed by contact of methane with the crust after it had solidified.) As a warm water–ammonia magma ascends it devolatilizes the clathrate, yielding methane fluid. However, diffusion arguments (Lunine and Stevenson, 1985) suggest this is unlikely, as only a thin rind of clathrate would be formed.

They suggest a layer of 10^{-6} m would form in 10^3 years, with the penetration depth $\sim \sqrt{\text{time}}$. Thus even after billions of years, a layer only $\sim 10^{-3}$ m thick should have formed. As the methane mass fraction of clathrate is < 0.14, the mass fraction of volatile (n_e) added to ascending magma (if the magma fills the crack) is $\sim 10^{-4}/d$ where d is the crack width, or $10^{-3} \sim 10^{-5}$.

As indicated above, these values are too low to significantly add to the volatile content of the magma.

Large volumes of gas could perhaps be liberated if some porous pocket of clathrate was met by ascending water–ammonia magma (Stevenson, 1982b; Lunine and Stevenson, 1987) but this is likely to lead to a pure "blowout" of the overburden, with little or no contribution of magma to the ejected clasts. Further, the existence of a porous clathrate pocket, without a similarly porous overburden, seems implausible.

These arguments apply only to the present atmosphere.

If in an earlier epoch, Titan's atmosphere was thinner, exsolution of dissolved volatiles (and perhaps the vapor pressure of the water–ammonia itself) may have been enough to disrupt the magma and lead to cryoclastic features like cinder cones.

Size of features

The size of features corresponding to single eruptive events may be limited by the volume of magma that can be accommodated in a crack. The maximum stable vertical length of a crack, according to Stevenson (1982a) is

$$Z < 2[K/\sqrt{\pi g(\Delta\rho)}]^{0.66} \qquad (6)$$

with K the stress intensity factor, estimated at $100 \, \text{MN m}^{-0.66}$ (Wadge, 1980) yielding $Z < 10$ km. The horizontal length L of the crack is not constrained, but most plausibly $> w$ and $< Z$. For a plausible maximum width of 1–10 m, the maximum volume per unit crack length of magma is $\sim 10^5 \, \text{m}^3 \, \text{m}^{-1}$.

Note that there is evidence of extremely large eruptive events on Miranda and Enceladus and elsewhere. This follows, as Stevenson (1982b) points out, from the inverse dependence on g which favors larger cracks (larger Z) on smaller (low-g) bodies.

To evaluate the width and thickness of a flow (or dome) with this volume, the "viscous drop" spreading model of Huppert et al. (1982) was applied, where the growth in drop radius r_N is described by

$$r_N = 0.894(g V^3 \rho/3\mu)^{0.125} t^{0.125}. \qquad (7)$$

This equation may be solved iteratively after estimating t, the spreading time of the drop, and using $V \sim \pi h_0 r_N^2$, with h_0 the height of the center of the drop.

It has been common, e.g. see McKenzie et al. (1992) and Jankowski and Squyres (1988), to use a conductive cooling time $\tau = h_0/\pi^2\kappa$ to estimate the spreading time of the drop. However, as discussed in Schenk (1991), a better estimate of the spreading time for basalt flows on Earth is given by a radiative cooling time $\Gamma = \rho C_p h_0/\epsilon\sigma T_0^3$, where ϵ is the emissivity (~ 1), σ the Stefan–Boltzmann constant ($5.67 \times 10^{-8} \, \text{W m}^{-2} \, \text{K}^{-4}$) and T_0 the eruption temperature.

Results are shown for $V = 10^5 \, \text{m}^3$ for various viscosities in Table 2, using both the conductive and radiative cooling times. The radiative cooling times, and hence the corresponding flow radii are probably strong upper limits, since (see later) the heat removal from a flow on Titan is much faster than by radiation alone: in this application, the conductive cooling time is perhaps a better estimate. Lava flows thinner than ~ 0.3 m seem improbable, so "flood" lavas seem unlikely on Titan.

(In the event that a ~ 10 km long crack on Titan propagated upwards, but erupted only through a small part of the crack, the effective magma volume per unit length of crack would be up to $10^8 \, \text{m}^3 \, \text{m}^{-1}$ and the radius of the drop would be ~ 20 times larger, and the height ~ 3 times larger: if it erupts along its whole length, the width and height of the flow would be similar to those shown, but obviously the length of the fissure would dominate the length of the lava flow.)

Table 2. Shapes of a $10^5\,\mathrm{m}^3$ cryolava flow on Titan

Viscosity (Pa s)	Conduction-limited cooling			Radiation-limited cooling		
	Radius (m)	Height (m)	Cooling time (s)	Radius (m)	Height (m)	Cooling time (s)
10	346	0.35	1.3×10^4	545	0.14	4.7×10^6
1000	236	0.76	5.8×10^4	344	0.36	1.2×10^6
10^5	160	1.65	2.8×10^5	217	0.90	3.0×10^6
10^7	109	3.5	1.2×10^6	137	2.25	7.1×10^6
10^9	75	7.6	5.8×10^6	86	5.67	1.9×10^7
10^{11}	51	16.3	2.7×10^7	54	14.3	4.7×10^7
10^{13}	34.6	35.3	1.2×10^8	34.4	35.8	1.2×10^8

Note: Effective viscosity used in computation is not necessarily true viscosity of fluid, as edges are chilled: effective viscosities are typically 4–6 orders of magnitude higher than the viscosity of the "pure" erupted liquid. Conduction-limited cooling times computed assuming $\kappa = 10^{-6}\,\mathrm{m^2\,s^{-1}}$, radiation-limited cooling times computed assuming $T = 176\,\mathrm{K}$ and emissivity $= 1$.

It may be seen that volcanic constructs should be relatively small on Titan, unless many eruptions took place in the same location.

Cooling rate of lavas—crust formation

When a magma erupts onto a planetary surface, it spreads and cools. The typically exponential increase in viscosity with falling temperature means that often the spread of a flow may be limited by the stiffness of a chilled crust, rather than the viscosity of the bulk lava.

Cooling occurs by radiation (the dominant cooling mechanism in terrestrial volcanism) and by convection. Following the methodology described in Griffiths and Fink (1992), I compute the cooling rates due to these two mechanisms for cryolavas on airless worlds (such as Enceladus), for Titan's present-day atmosphere, and in a liquid hydrocarbon sea on Titan. It may be seen that lavas cool ~ 2 orders of magnitude faster on Titan than if they cooled by radiation alone (since cryolavas erupt at temperatures an order of magnitude lower than silicate lavas, radiative cooling may be expected to be much less important, so convective effects dominate). Further, eruption into a liquid hydrocarbon is another two orders of magnitude faster.

For comparison, I show (Fig. 2) the cooling rate of silicate lavas on the Earth, Moon, Venus and for submarine volcanism on Earth. The cooling curves for the Moon and Earth are similar, showing that convective cooling is relatively unimportant for hot silicate lavas, although the thick atmosphere of Venus is somewhat more efficient at cooling. (Due to the high surface temperature on Venus, the overall cooling rate slows to lower than that on Earth, after $\sim 10^5\,\mathrm{s}$.) As might be expected, cooling in the ocean is fast.

Griffiths and Fink suggest that the characteristic time t_c for a crust to form is the time taken for the surface of the flow to cool to the glass transition temperature. (As these cooling times are usually short, latent heat effects are ignored as a glass typically forms, with crystallization occurring much later. As the glass transition of water–ammonia is poorly-characterized at present, I use the freezing point of the peritectic.)

From Fig. 3, we see $t_c \sim 10^3\,\mathrm{s}$ for eruption onto Titan with cooling by the atmosphere ($\sim 10^5\,\mathrm{s}$ for radiative cooling only). Eruptions into a lake or sea on Titan have $t_c \sim 0.2\,\mathrm{s}$, however.

The morphology of a lava flow may be characterized by the ratio Ψ of this cooling time to an advection time t_a, defined by

$$\Psi = t_c/t_a, \text{ with } t_a = (\mu/g\rho)^{0.66}q^{-0.33} \qquad (8)$$

and $q = Q/L$. As described in Griffiths and Fink (1992) the lava morphology can be crustless, with perhaps levees ($\Psi > 18$), folded ($18 > \Psi > 10$), rifted ($10 > \Psi > 3$), or pillowed ($\Psi < 3$). For water–ammonia slurry ($\mu \sim 10^3\,\mathrm{Pa\,s}$), $t_a \sim 64$–$1.5\,\mathrm{s}$ for the eruption rates given in Table 1, hence $\Psi = 15$–680, giving at most some slight folding of the lava for subaerial land eruptions on Titan. For crystal-free water–ammonia ($\mu \sim 10\,\mathrm{Pa\,s}$), $\Psi \sim 10^3$–$10^5\,\mathrm{s}$, giving fairly smooth lava flows, somewhat akin to basalt flooding (although see the earlier section for a probable lower limit on flow thickness).

For both of these fluids, however, eruption into a liquid hydrocarbon gives $\Psi < 1$, suggesting that pillow lava will form. (In principle, high-viscosity lavas could lead to $\Psi < 1$ for subaerial eruptions, but the high viscosities make magma ascent difficult—see Fig. 1.)

Observation of pillow lava formations on Titan's surface would therefore imply that these regions had been submerged when the eruption occurred. Thus constraints might be placed on the depth history of hydrocarbons on Titan's surface.

Note that although convection plays a major role, the influence of a thicker atmosphere (such as a 5 bar atmosphere at 125 K (McKay et al., 1993)) is modest (and the lava cooling curve for such an atmosphere cannot be discriminated from that for Titan's present atmosphere). Thus pillow lavas should form only in undersea eruptions on Titan.

After the formation of the crust, the cooling of a lava flow is usually limited by conduction through the body of the lava itself. (Note a lava flow also loses heat by conduction through its base, but this is less important for considering the lava flow morphology than the convective cooling mechanism.) This forms the principal justification for using the conductive cooling time in the previous sec-
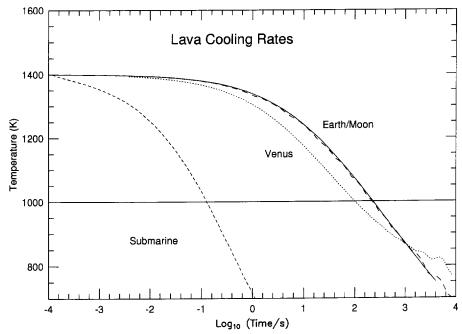
Lava Cooling Rates

Fig. 2. Cooling rates for basalt lavas on the terrestrial planets. Crust formation (the time taken for the curve to reach the glass transition temperature) takes $\sim 10^2$ s, but < 1 s for submarine eruptions, due to rapid convective cooling. The effect of gas convective cooling is very small, although note the slow late cooling in Venus' hot atmosphere

tion: note, however, that the growth of a thick crust will certainly modify the effective viscosity of the flow, and may even cause a lava flow model based on a rigid crust (Iversen, 1987) to be more applicable. However, in the study of dome volcanoes on Venus by McKenzie *et al.* (1992), the Huppert viscous drop (Huppert, 1982) model was found to yield better fits to the shapes of observed domes.

The "pillows" in a submarine flow are typically of the

order of 1 m across (Chadwick and Embley, 1994). As only a tiny part of Titan's surface will be imaged at this resolution (Lorenz, 1994), it seems unlikely that we would see an unmistakeable pillow lava. However, synthetic radar mapping from orbit will cover $\sim 20\%$ of Titan's surface, and it may be possible to derive textural information (and hence infer pillow-type lavas) from this data, albeit at lower spatial resolution (~ 500 m, of the order of the larger flows suggested here).
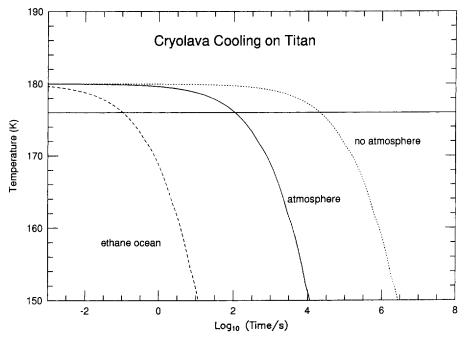
Cryolava Cooling on Titan

Fig. 3. Corresponding cooling curves for water–ammonia cryomagmas: the influence of an atmosphere is remarkable, as radiative cooling at these low temperatures is inefficient. Quenching in liquid hydrocarbon oceans is very rapid, and comparable to cooling in terrestrial submarine volcanism

Resurfacing rate

Some (very broad) constraints may be placed on the eruption or resurfacing rate on Titan. Cassini measurements (notably radar sensing of crater populations—Lorenz (1995)) will help constrain the resurfacing history. Additionally, measurement of the argon abundance may help constrain the amount of volcanism in Titan's past (Engel et al., 1994).

First, let us assume Titan's global heat flow of $\sim 4 \times 10^{11}\,W$ is transferred to the surface by the ascent of magma. Assuming the latent heat of fusion of water–ammonia at $\sim 100\,J\,g^{-1}$, then the heat flow allows $\sim 5 \times 10^6\,kg\,s^{-1}$ of magma, or $\sim 1.5 \times 10^{14}\,kg\,yr^{-1}$. Over Titan's surface area of $\sim 8 \times 10^{13}\,m^2$, this yields an upper bound of $\sim 2\,kg\,m^{-2}\,yr^{-1}$, or $\sim 0.02\,m\,yr^{-1}$. Since on the Earth only $\sim 10\%$ of the geothermal heat flow is used in volcanism, erupting $\sim 20\,km^3$ of basalt per year (Francis, 1993), a more reasonable upper limit for Titan's resurfacing rate is $0.002\,m\,yr^{-1}$.

A second indirect constraint may be deduced from the methane content of water–ammonia eruptions. Methane photolysis (Yung et al., 1984; Lara et al., 1994) in the atmosphere would have destroyed of the order of 500 m depth of liquid methane over the age of the solar system (4.5 Gyr), thus the loss rate is $\sim 2 \times 10^{-12}\,kg\,m^{-2}\,s^{-1}$, or $\sim 7 \times 10^{-5}\,kg\,m^{-2}\,yr^{-1}$.

If, on the other hand the present-day photolysis rate is being compensated by a resupply of methane by volcanism, rather than a surface reservoir, then if the methane makes up less than 5×10^{-3} of the mass of a water–ammonia eruption, then the above methane flux corresponds to a cryolava eruption rate of $> 0.014\,kg\,m^{-2}\,yr^{-1}$. Thus, there exists a present-day range of resurfacing rate between 0.014 and $2\,kg\,m^{-2}\,yr^{-1}$ which is compatible with the two constraints: continuous eruption is a viable means of methane resupply. On the other hand, eruption rates could be rather lower (or, indeed, zero) if methane was erupted in the past when the heat flow was higher, and is either ephemeral (to be depleted by photolysis in $\sim 10^7$ years) or is buffered by a surface or porous near-surface reservoir. Lower eruption rates would also be possible if the methane content of cryo-lavas is higher than the 5×10^{-3} value I have cited.

Conclusions and predictions

Present-day cryovolcanic activity seems possible, with magma transport by cracks to the surface. Heat flow and methane photolysis constraints lead to a suggested resurfacing rate of 1.4×10^{-5}–$2 \times 10^{-3}\,m\,yr^{-1}$ for steady-state conditions.

Any eruptions today or in the recent past are likely to have been modest, and either effusive, forming small flows or domes (a few hundreds of meters across, at most), or less likely, explosive, resulting from blowout of moderately deep volatile reservoirs. Cryoclastic eruptions seem improbable: if Cassini reveals "cinder cones", we might expect to have to revise either our understanding of cryovolcanism on Titan, or deduce that Titan's atmospheric

pressure has been much lower when the eruption occurred (a possibility that could be verified by examining the cratering population on Titan—Engel et al. (1995)).

Eruptions under present-day or ancient seas or lakes may look like terrestrial "pillow" lavas, owing to the high rate of cooling by liquid hydrocarbons. Eruptions into any atmosphere (except at extremely low eruption rates) are unlikely to look this way.

Crater basins seem likely places to find pillow lavas, as they penetrate below the regolith, crustal stresses are likely to exist around craters, and craters are likely to have held accumulations of liquids.

The coverage of Titan by Cassini instrumentation, however, is such that we will be lucky to recognize such pillow lavas. Reconaissance at the required resolution must await future missions.

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