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Silicate Volcanism on Io

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Io is currently emitting 1–1.5 W m^{-2} of tidal energy as a result of its volcanic activity. If the lithosphere is more than 20 km thick, as appears probable from the surface relief, only a fraction of the tidal energy can be dissipated within the lithosphere, otherwise it will become thinner. The rest of the tidal energy must be dissipated below the lithosphere. Io is likely to be highly differentiated as a result of the volcanic activity, with a low melting temperature fraction, such as basalt, near the surface and a high melting temperature fraction, such as peridotite, at depth. If solidus temperatures are reached at depths shallower than the thickness of the basalt layer, a partial melt zone will separate an all-basalt lithosphere above from a peridotite mantle below, and part of the tidal energy will be dissipated by viscous deformation within the melt zone. Most of the energy dissipated below the lithosphere will be transported upward through the lithosphere as silicate magma, which is generated in quantities sufficient to resurface the satellite at a rate of a few tenths of a centimeter a year, depending on the lithosphere thickness. Many of the characteristics of the Ionian surface have been explained in terms of sulfur volcanism. However, most of the features observed can be as readily explained by silicate volcanism, and silicates are more consistent with the apparent strength of the surface as implied by the relief. Simulations of basaltic eruptions indicate that the surface temperatures that would result from basaltic eruptions are similar to those measured by the Voyager infrared interferometer spectrometer experiment. The high rates of emission by the Ionian hot spots imply eruption rates that are high compared with typical terrestrial eruptions. An eruption rate of 4000 m^3 s^{-1} may be required to explain Loki which is currently emitting 10^{13} W. Although the near-surface materials are mainly silicate, they may contain several percent of volatile components rich in S, Na, and K. Remobilization of these components by the ongoing silicate eruptions causes the plumes, provides material to the torus, and gives the satellite its characteristic reflectivity.

INTRODUCTION

The main question that this paper addresses is whether volcanic eruptions on Io are mainly of silicates or sulfur. Following the Voyager encounters with Jupiter, two views developed concerning the nature of Io's surface (see *Kieffer* [1982] for discussion). One view was that the surface of Io is covered with several kilometers of sulfur and that the volcanism is predominantly, if not exlusively, of sulfur-rich compounds [*Smith et al.*, 1979*b*]. The other view was that both silicate and sulfur volcanism occurred and that the near-surface consists of interbedded silicate and sulfur deposits [*Carr et al.*, 1979]. The sulfur-rich model was proposed for a variety of reasons: SO_2 had been positively identified from the IR spectrum [*Fanale et al.*, 1979; *Smythe et al.*, 1979]; a strong UV absorption feature in Io's reflection spectrum was possibly explained by sulfur [*Nash and Fanale*, 1977; *Nelson and Hapke*, 1978]; different sulfur allotropes appeared to match the different colors on Io's surface [*Sagan*, 1979]; black spots within calderas were thought to have a similar albedo to liquid sulfur [*Smith et al.*, 1979*a*]; temperatures measured at hot spots ranged up to 600°K, consistent with sulfur volcanism; finally, ionized sulfur was detected in Io's torus [*Kupo et al.*, 1976], and Io was an obvious source. These observations led *Smith et al.* [1979*b*] to conclude that "the upper crust consists largely of elemental sulfur and SO_2 and overlies a subjacent layer of molten sulfur possibly several kilometers thick." They also interpreted "all the black features resembling lava lakes" as "active or recently active lakes of molten sulfur." Smith et al. raised two main arguments against any silicate volcanism. First, they claimed that silicates are not buoyant enough to rise through the sulfur-rich materials near the surface. Second, they suggested that if silicate magmas did get close to the surface, they would irreversibly remove the sulfur. Sulfur would react with the lava, then burial and melting would result in segregation of sulfur-rich melts which would accumulate permanently at the base of the lithosphere. Because we have clear evidence of sulfur at the surface, they claimed that silicate volcanism is insignificant.

Since the Voyager encounters, hybrid models in which both silicates and sulfur participate in the volcanism have gained favor [*Clow and Carr*, 1980; *Schaber*, 1982; *McEwan and Soderblom*, 1983; *Lunine and Stephenson*, 1986]. However, while few workers now believe in the sulfur ocean model, the role of sulfur in the evolution of the surface remains controversial. The dark flows on Io's surface are widely interpreted as the result of sulfur eruptions [*Pieri et al.*, 1984; *Fink et al.*, 1983; *Sagan*, 1979; S. M. Baloga et al., unpublished manuscript, 1986] despite the arguments of *Young* [1984] that there are no sulfur flows and despite the arguments of *Clow and Carr* [1980] that the strength of the surface is inconsistent with it being dominantly sulfur. Furthermore, although the dark floors of calderas are still widely interpreted as sulfur lava lakes [*McEwan et al.*, 1985; *Pearl and Sinton*, 1982; *Lunine and Stevenson*, 1985], it will be argued here that the dark floors could equally, if not more probably, be interpreted as the result of basaltic eruptions.

This paper reexamines several aspects of silicate volcanism on Io, such as how and in what quantities silicate magma could be generated, how it might reach the surface, and how it might manifest itself there. Of special interest are the surface temperatures that would result as a consequence of silicate eruptions, since the temperatures measured have been regarded as strong evidence against silicate eruptions and strong support for the presence of low-temperature melts at the surface. The general thrust of the paper is that silicates are the principal means by which the satellite dissipates its heat and that they play a prominant role in the evolution of the surface. They not only continually remobilize the more volatile components of the surface, such as sulfur, but they also are erupted onto the surface and form many, if not most, of the landforms. The present configuration of the Ionian surface is thus

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the result of both silicate and sulfur eruptions. Sulfur compounds cause most of the tinting of the surface; silicates constitute the bulk of the surface material, thereby providing the surface with the strength to support the observed relief.

GENERATION OF SILICATE MAGMA

According to the tidal theory of heat generation within Io, as originally proposed by *Peale et al.* [1979], all the tidal energy is presently dissipated within a thin, rigid lithosphere that overlies a completely molten interior. Estimates of the thickness of the rigid rind range from 8 to 18 km [*Peale et al.*, 1979; *Nash et al.*, 1985]. If all the tidal energy were dissipated uniformly within a thin, rigid, homogeneous lithosphere, in which the temperature profile had equilibrated with the energy generated, no silicate volcanism would occur. All the tidal energy would be deposited at depths shallower than the zone of melting and would be conducted upward toward the surface. The energy would be available to drive volcanic processes involving volatile species such as sulfur and sulfur dioxide, but silicate melts would result only if there were major anisotropies which both localized tidal dissipation and hindered conduction away from the dissipation sites.

A lithosphere thin enough to conduct away all the tidal energy dissipated appears inconsistent with the presence of 10-km-high mountains [*Smith et al.*, 1979*a*] and 2- to 3-km-deep calderas [*Clow and Carr*, 1980; *Schaber*, 1982] and with localization of thermal emission in discrete hot spots. Although the thickness of the lithosphere has not been precisely defined, *Nash et al.* [1985] estimate that it must be at least 30 km thick in the regions where the 10-km-high mountains occur. The estimate is derived by assuming isostasy and that the density difference between the mountains and the plains is no larger than 0.5 g cm^{-3}. The lithosphere is probably not the same thickness everywhere. The mountains, along with other features such as plumes, are distributed nonuniformly [*McEwan and Soderblom*, 1983], as is current volcanic activity as indicated by the pattern of thermal emission [*Johnson et al.*, 1984]. We should accordingly expect the lithosphere thickness to be nonuniform. Calderas suggest, however, that the lithosphere is thick even away from the mountains. Calderas several tends of kilometers across and hundreds of meters to kilometers deep are seen in all near-terminator pictures [*Schaber*, 1982; *Clow and Carr*, 1980], and there is no reason to assume that the areas that were near the terminator when the Voyager spacecraft flew by are atypical. Calderas form by faulting following removal of magma from a chamber within the lithosphere. Where identifiable under terrestrial calderas, the magma chamber is comparable in diameter to the resulting caldera [*Wood*, 1984]. Thus the Ionian lithosphere is likely to be everywhere thick enough to contain magma chambers several tens of kilometers across, and the chambers must be buried sufficiently deeply that following evacuation of magma the roof collapses as a coherent unit several tens of kilometers across rather than breaking apart. Unfortunately, no quantitative estimates have been made as to how thick the lithosphere must be to allow formation of the observed calderas, but it appears unlikely that calderas several tens of kilometers across could form if the lithosphere thickness is close to the estimated minimum of 8–18 km. The pattern of emitted radiation also suggests lithosphere that is not so thin as to allow all the tidal energy to be lost through conduction. *Matson et al.* [1981] demonstrate that most of the tidal energy is emitted at hot spots and that only a fraction of the tidal energy is lost through conduction. This implies, first, that the lithosphere is thicker than the minimum value of 8–18 km and, second, that a significant fraction of the tidal heat generated is transported toward the surface as magma. Thus several lines of evidence suggest that the lithosphere is too thick to be able conduct all the tidal energy to the surface and that some of the energy is carried to the surface as silicate magma.

If a significant fraction of the tidal energy dissipated is transported to the surface as silicate magma, then all the tidal energy cannot be dissipated in the rigid lithosphere as originally suggested by *Peale et al.* [1979]; some must be dissipated below the rigid lithosphere. Several investigators have proposed modifications to the Peale et al. model. *Schubert et al.* [1981] and *Cassen et al.* [1982] suggest that the outer skin could be underlain by a partially molten zone, which is in turn underlain by an essentially rigid interior. In these models, tidal heating still occurs in the outer skin. C. F. Yoder and J. Faulkner (manuscript in preparation, 1986) also examine a model in which the rigid crust is decoupled from the solid interior by a melt zone but they suggest that a significant fraction of the tidal energy could be dissipated by skin friction at the boundaries of the melt zone. *Ross and Schubert* [1984] alternatively proposed that a thin rigid skin could overlie a partially molten interior, as originally suggested by *Peale et al.* [1979], but they propose that as much, if not more, of the tidal energy is dissipated by viscous deformation in the interior as is generated by flexing of the rigid shell. In this paper the amount of tidal energy that can be dissipated in the lithosphere is assessed by determining the range of temperature profiles possible within a lithosphere of a given thickness. One conclusion reached is that all of the tidal energy is unlikely to be dissipated in a thin rigid shell; some must be dissipated by viscous heating below the lithosphere.

The rate of magma generation in Io depends on the magnitude and distribution of heat sources. Various estimates of the sources of heat within Io indicate that tidal heating greatly predominates. Radioactive heating is estimated to generate 0.011–0.015 W m^{-2} of the surface heat flux depending on whether it is lunar or chondritic in composition [*Cassen et al.*, 1982]. The maximum electrical heating that can result from the passage of Io through Jupiter's magnetic field is only about 0.005 W m^{-2} [*Goldreich and Lynden-Bell*, 1969; *Colburn*, 1980]. In contrast, heat losses observed are in the range of 1–1.3 W m^{-2} [*Johnson et al.*, 1984], and tidal heating is the only plausible explanation so far proposed for heat generation of this magnitude. In recent years a discrepancy arose between the heat fluxes estimated from infrared observations [*Matson et al.*, 1981; *Sinton*, 1981; *Pearl and Sinton*, 1982; *Morrison and Telesco*, 1980], which ranged from 1.5 to 2 W m^{-2} and an upper theoretical limit of 1 W m^{-2} imposed by the lower limit of 6×10^4 on the Q of Jupiter [*Nash et al.*, 1985]. Recently, this apparent discrepancy has been partly resolved. The infrared-based estimates of global heat flow assumed global uniformity in infrared emission. *Johnson et al.* [1984] have shown that almost all of Io's emission is through a few volcanic hot spots, which are distributed nonuniformly in longitude. The early telescopic infrared measurements were made during eclipses and were restricted by the geometry to longitudes primarily between 270°W and 360°W. Johnson et al. showed that emission from these longitudes is anomalously high, partly because of inclusion of the very active volcano Loki. Global estimates of emission based on longitudinal uniformity were therefore also high. The revised estimate by Johnson et al. of the emitted power is 1.0–1.3 W m^{-2}, which is still somewhat higher than the limit imposed by the present Q of Jupiter. In the subsequent discussion the power emitted is assumed to be around 1 W m^{-2}, close to the upper bound based on Jupiter's Q and the lower bound based on the infrared observations.

Temperature profiles within the Ionian lithosphere were calculated for various limiting assumptions and various lithosphere thicknesses. The main purpose was to determine the maximum amount of heat that could be generated in a lithosphere of a given thickness without causing it to thin. The deficit with respect to the total of 1 W m^{-2} must then be dissipated below the lithosphere and be available for generation of silicate magma. In the simplest possible model, no heat is generated within the lithosphere, and the thermal conductivity is uniform. Conductive heat loss is then simply $K\Delta T/L$, where K is the thermal conductivity, ΔT is the difference between the solidus and surface temperatures, and L is the lithosphere thickness. Taking values of 4.5 W m^{-1} $°K^{-1}$ for K, 1300°K for the solidus temperature, and 100°K for the surface temperature gives a conductive heat loss of 0.18 W m^{-2} at the surface for a 30-km-thick lithosphere. Such a simple model is, however, unrealistic on several grounds but mainly because it assumes that all the heat is generated below the lithosphere and none by tidal flexing of the lithosphere itself.

A more probable model is one in which the thickness of the lithosphere is controlled by the heat tidally generated within it. For maximum tidal dissipation within the lithosphere, no heat is conducted into it from below, the temperature profile is in conductive equilibrium with the heat generated, and temperatures reach the solidus at the base of the lithosphere. When these conditions are achieved, any increase in the rate energy input will cause the lithosphere to thin. Temperature profiles within the lithosphere were calculated numerically so that different thermal conductivities could be assigned to different parts of the profile. The lithosphere was divided into 100–200 cells, appropriate conductivities were assigned to each cell as discussed below, and a surface temperature of 100°K was assumed. For a given lithosphere thickness, heat production rates per unit volume were sought that produced equilibrium temperature profiles satisfying the constraint that solidus temperatures are reached at the base of the lithosphere. Heat production was assumed to be uniform throughout the profile. After the profile converged, the total energy that could be dissipated in the lithosphere was derived from the heat production per unit volume.

In the simplest of such profiles, all the heat generated within the lithosphere is assumed to be conducted to the surface, and the entire lithosphere is assumed to have the conductivity of silicates. *Toksoz and Johnston* [1974] in reviewing what conductivity is appropriate for modeling planetary interiors found so much uncertainty in different estimates that they concluded that a constant conductivity was as good an approximation of the available experimental data as any temperature dependent model. To test the sensitivity of heat fluxes to conductivity, profiles were calculated both by using a constant K of 4.5 W m^{-1} $°K^{-1}$, the value prefered by Toksoz and Johnston, and by using the variable K suggested by *Schatz and Simmons* [1972]. In the latter model, $K = K_L + K_r$, where K_L is the lattice conductivity equal to $418.4/(30.6 \times 0.12T)$ and K_r is the radiative conductivity equal to 0 for T less than 500°K and $230(T - 500)/10^7$ for T greater than 500°K. Units are watts per meter per degree kelvin and T is absolute temperature. For a 30-km lithosphere with a constant thermal conductivity of 4.5 W m^{-1} $°K^{-1}$ the maximum energy than can be deposited within the lithosphere without causing it to thin is 1.2×10^{-5} W m^{-3}, giving a flux at the surface of 0.36 W m^{-2}. For the Schatz and Simmons conductivity the surface flux is 0.28 W m^{-2}. These values scale inversely with lithosphere thickness, so for a 60-km-thick lithosphere the values are 0.18 and 0.14 W m^{-2}, respectively, and for a 10-km lithosphere the values are 1.2 and 0.9 W m^{-2}. In the subsequent calculations a constant silicate conductivity was assumed; the assumption is conservative in the sense that it maximizes the amount of energy that can be dissipated in the lithosphere.

The profiles just discussed assume that all the heat generated within the lithosphere is conducted outward at a rate controlled by the conductivity of silicates. Such an assumption is reasonable for the intermediate and deeper parts of the profile because the heat generated within the lithosphere cannot be convected to the surface as silicate magma once the lithosphere has stabilized unless there are gross anisotropies that cause localization of heat dissipation. However, presence of sulfur and mobilization of sulfur-rich melts could alter the temperature profile near the surface. Possible effects of sulfur on the temperature profiles were accordingly examined, despite scepticism that bodies of sulfur large enough to significantly perturb the temperature profiles throughout the lithosphere exist in Io's crust. Sulfur could affect the profiles in two opposing ways. First, the presence of significant amounts of sulfur near the surface could give the near-surface materials a lower conductivity than silicates, thereby steepening the near-surface temperature gradient. Alternatively, transport of sulfur magma to the surface could increase the effective conductivity, thereby making the near-surface temperature gradient shallower.

In the first case, where the insulating effects of sulfur dominate over convection of sulfur-rich magma, an upper limit can be placed on the near-surface temperature gradient. If several percent sulfur is present, the gradient cannot be so steep that sulfur melts at depths less then 2 km; otherwise the observed calderas would collapse or fill with molten sulfur [*Clow and Carr*, 1980]. Profiles were therefore calculated with the temperature at 2 km depth equal to the melting temperature of sulfur, 393°K. With these assumptions, no more than 0.31 W m^{-2} can be dissipated in a 30-km-thick lithosphere, and no more than 0.18 W m^{-2} can be dissipated in a 60-km-thick lithosphere (Figures 1 and 2). The temperature profile is less constrained if it is controlled by transport of sulfur-rich magmas to the surface. If sulfur flows are erupted onto the surface, then sulfur melting temperatures must be reached at the base of some sulfur-rich layer. Assuming somewhat arbitrarily an extreme model in which Io is covered with a 5-km-thick layer which is rich in sulfur and sulfur is at its melting temperature at the base of the layer, then a 30-km-thick lithosphere could dissipate 0.41 W m^{-2} (Figures 1 and 2). The equivalent figure for a 60-km-thick lithosphere is 0.17 W m^{-2}. These figures are close to the all-silicate model described above. If the 5-km-thick layer had the thermal conductivity of sulfur, then the conductive heat loss at the surface would be only 0.025 W m^{-2} or 2.5% of the total heat lost by the satellite, the rest being lost by transport of magma to the surface. The estimates of the maximum amount of heat that can be dissipated within lithospheres of different thicknesses, summarized in Table 1, are relatively insensitive to the amounts of sulfur close to the surface. Rates of generation of silicate magma determined below from these estimates are therefore consistent with both sulfur-rich and sulfur-poor models of the surface.

Thus some rather simple geologic reasoning leads to limits on the amount of tidal energy that can be dissipated in the lithosphere as a function of its average thickness. The difference between the total energy tidally dissipated in the satellite (1 W m^{-2}) and that which can be dissipated in the lithosphere gives the amount that must be released below the lithosphere. With maximum energy dissipation in the lithosphere, no heat is conducted into the lithosphere from below; all the energy

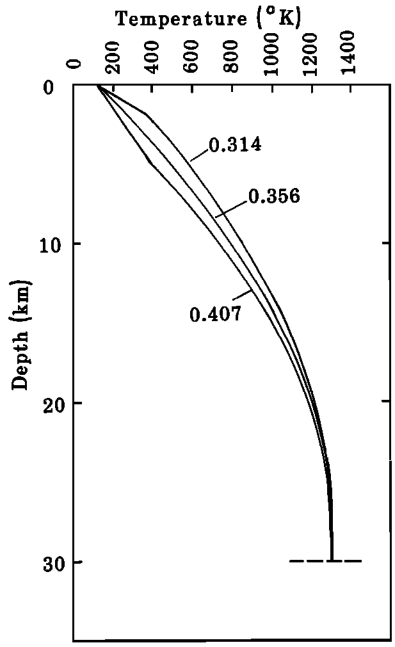

Fig. 1. Possible temperature profiles in a 30-km-thick lithosphere. Middle curve assumes a conductivity of silicates. The upper curve assumes that sulfur is at its melting temperature (393°K) at a depth of 2 km and silicates occur below that depth. The lower curve assumes a sulfur melting temperature at 5 km and silicates below. Also indicated are the rates of heat generation within the lithosphere (in watts per square meter of the surface) that are needed to maintain the profiles.

released below the lithosphere goes into formation of silicate melt. For any given thickness of lithosphere we can therefore calculate the minimum rate of magma generation from the profiles shown in Figures 1 and 2 and their equivalents. The generation rate is a minimum since the profiles are calculated for maximum heat dissipation within the lithosphere, which need not necessarily be achieved. The maximum rate of magma generation for a given lithosphere thickness is calculated on the assumption that all the tidal energy is dissipated deep within the planet and none in the lithosphere. In this case the maximum amount of energy is conducted up through the base of the lithosphere, and the remainder is used to generate silicate magma. Figure 3 shows the rate of generation of silicate magma for different lithosphere thicknesses based on these two limiting assumptions. If, as seems likely, the present structure of the satellite is in equilibrium with the present thermal regime, then the lithosphere is getting neither thinner nor thicker, and magma is transported upward out of the melt zone at the same rate that it is generated. In Figure 3 this heat loss is indicated in terms of a resurfacing rate following *Reynolds et al.* [1980], who showed that a silicate resurfacing rate rate of 1 cm yr^{-1} results in removal of heat at a rate of 1.8 W m^{-2}. The fate of the magma as it works its way up to the surface will be discussed in a later section.

VISCOUS DISSIPATION IN THE MELT ZONE

The above discussion and Figures 1 and 2 show that at least 50% of the tidal energy is dissipated below the lithosphere if it is 30 km thick and at least 80% if it is 60 km thick. Precisely how this dissipation occurs is unclear. One possibility is that Io is highly differentiated and has a crust of a low melting temperature fraction such as basalt, which overlies a higher melting temperature fraction such a peridotite. If basalt solidus temperatures are reached at depths that are shallower than the thickness of the basalt layer, the lithosphere would be all basalt and be underlain by partially melted basalt. Beneath the melt zone would be solid peridotite. This model is equivalent to model III of *Schubert et al.* [1981]. The viscosity of basalt drops from around 10^{20} P, near its solidus at 1250°K, to around 10^4 P at 1400°K, where it is 70–80% melt [*Shaw*, 1969]. The drop is about an order of magnitude for every 10°K increase in temperature. Once melting starts, therefore, the rigid lithosphere will tend to be decoupled from the interior by a zone which deforms viscously. Shear between the interior and the lithosphere, caused by differences in the amplitude and phase of their tidal deformations, will be accommodated by viscous deformation in the melt zone, thereby causing heating and a further lowering of the viscosity. The process will continue until a balance is reached between the rate of melting and the rate of transport of magma upward out of the melt zone.

Shaw [1970] proposed such a model for the earth. He suggested that viscous dissipation of tidal energy could cause shear melting in the upper mantle. Following *Zener* [1948], he suggested that such dissipation would be maximized when the product of the angular frequency of deformation and the relaxation time is unity. For Io the tidal frequency is 4.1 $\times 10^{-5}$ rad s^{-1} giving a relaxation time τ of 2.4×10^4 s at maximimum dissipative efficiency. For the Maxwell model of viscoelastic behavior $\tau = \eta/M$ where η is the apparent viscosity and M is the shear modulus [*Van Wazer et al.*, 1963]. Assuming a typical shear modulus of 3×10^{11} dyn cm^{-2} gives 7×10^{15} P for the viscosity of the melt zone, which is the value for basalt containing 10–25% melt [*Shaw*, 1969]. Some of the tidal energy could therefore be dissipated in a zone of partial melt at the base of the lithosphere. As the energy is dissipated, more melt will form and the interstitial pressure will rise, a consequence of the increase in volume on melting. Ultimately, the pressure rise and the density difference between the magma and the overlying rocks will force the magma out of the melt zone toward the surface. The volume lost will be replaced by cooler materials from above, and the cycle will start again. *Ross and Shubert* [1984] alternatively proposed a two-layer model in which the rigid lithosphere

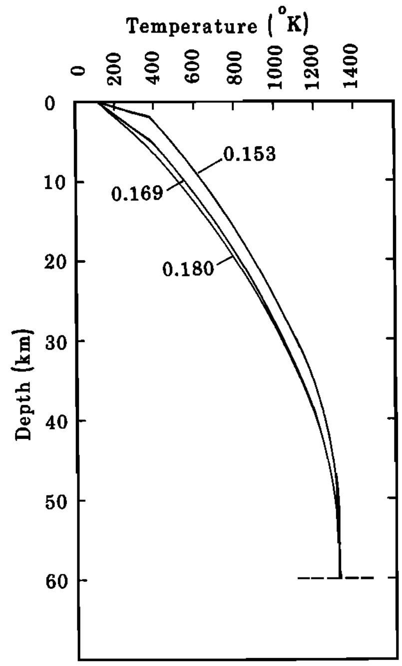

Fig. 2. Possible temperature profiles in a 60-km-thick lithosphere. Lower curve assumes the conductivity of silicates. The lower and upper of the two remaining curves assume 393°K temperatures at 5 and 2 km depths, respectively, as in Figure 1.

TABLE 1. Maximum Heat That Can Be Dissipated Within 30-km and 60-km Lithospheres Under Different Assumptions

Assumption	30 km	60 km
All silicate, conductivity 4.5 W m^{-1} °K^{-1}	0.36	0.18
All silicate, Schatz and Simmonds conductivity	0.28	0.14
T = 393°K at 2 km, conductivity 4.5 W m^{-1} °K^{-1} below	0.31	0.18
T = 393°K at 5 km, conductivity 4.5 W m^{-1} °K^{-1} below	0.41	0.17

Heat values are in watts per square meter of the surface.

overlies an interior that deforms viscously. They suggest that viscous dissipation could occur throughout the interior of the body, not just in a layer at the base of the lithosphere, and estimate an interior viscosity of 10^{11} P, corresponding to a 60% melt.

Whichever the mechanism, it appears likely that (1) if the rigid lithosphere is more than 20 km thick, as is probable both from the surface relief and from the large fraction of the satellites energy that is radiated from hot spots, then a significant fraction of the tidal energy is dissipated in viscous heating, and (2) most of this heat is used in generating silicate magma which is ultimately transported upward out of the melt zone.

THERMAL ANOMALIES AT ERUPTION SITES

Given that silicate magma is transported up through the Ionian lithosphere in order to dispose of much of the tidal energy, the question arises as to whether the magma mostly reaches the surface to form volcanic landforms, or whether it is mostly intruded into the lithosphere at relatively shallow depths. Apart from the plumes, the most direct evidence of volcanic activity is the presence of thermal anomalies or hot spots [*Hanel et al.*, 1979; *Pearl and Sinton*, 1982]. *Morrison and Telesco* [1980] modeled the nonsolar part of Io's infrared eclipse spectrum on the basis of three components. They suggested that 9.4×10^{-6} of the surface area is at 600°K, 2.2×10^{-4} at 350°K, and 1.4×10^{-2} at 200°K. *Sinton* [1981] found a good fit with 2.1×10^{-5} of the area at 615°K and 3.9×10^{-3} of the area at 294°K. *Pearl and Sinton* [1982] recognized three types of hot spots from the Voyager infrared interferometer spectrometer (IRIS) data: stable, high-temperature (~600°K) sources; transient, high-temperature (~600°K) sources; and stable, low-temperature (<400°K) sources. Most of the anomalies are associated with local dark areas, and in general, the darker the area the higher the temperature [*McEwan*, 1984]. Where the resolution is good enough the dark spots are observed to lie mostly within calderas. The favored interpretation of the dark hot spots is that they are lakes of molten sulfur, the result either of eruption of sulfur magma from kilometer depths [*Smith et al.*, 1979*a*, *b*; *Sagan*, 1979] or mobilization of sulfur near the surface by silicate intrusions [*Soderblom et al.*, 1980; *McEwan and Soderblom*, 1983; *Lunine and Stevenson*, 1986]. The sulfur lake interpretation is based on the temperatures which are believed to be too low for silicates, on the similarity in albedo between the dark spots and liquid sulfur, and on the abundant evidence, outlined at the beginning of this paper, that sulfur is abundant on the Ionian surface. Implicit in the rejection of a silicate explanation of the anomalies is a belief that many of the dark areas are active lava lakes, so that the observed low temperatures imply low-temperature melts. It will be demonstrated below, however, that the temperature measurements are close to those expected from prolonged eruption of basaltic lavas. Boiling sulfur lakes hundreds of kilometers in diameter are not required to explain them. (The dark horseshoe-shaped "lake" of Loki covers an area 2.5 times the area of the island of Hawaii; the IRIS field of view of Loki covers an area of 378,000 km^2, 36 times the area of Hawaii (Figure 4)).

Before exploring what temperatures are to be expected from

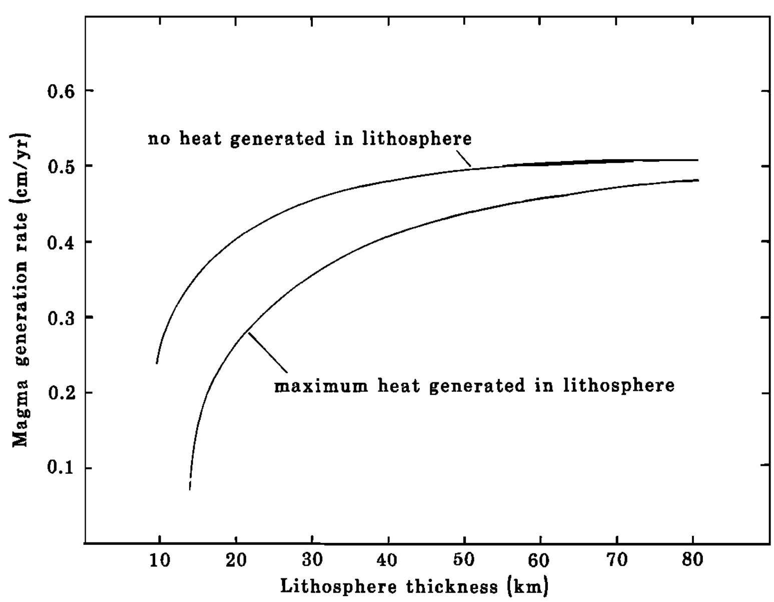

Fig. 3. Maximum and minimum rates of magma generation as a function of lithosphere thickness. The amounts of magma generated are expressed as resurfacing rates. The upper curve assumes no heat generated in the lithosphere, so that the heat available for melting is the total tidal heat generated (1 W m^{-2}) less the conductive heat loss. The lower curve is derived from the difference between 1 W m^{-2} and the maximum amount of heat that can be generated in the lithosphere as indicated by profiles similar to those in Figures 1 and 2 for different thicknesses.

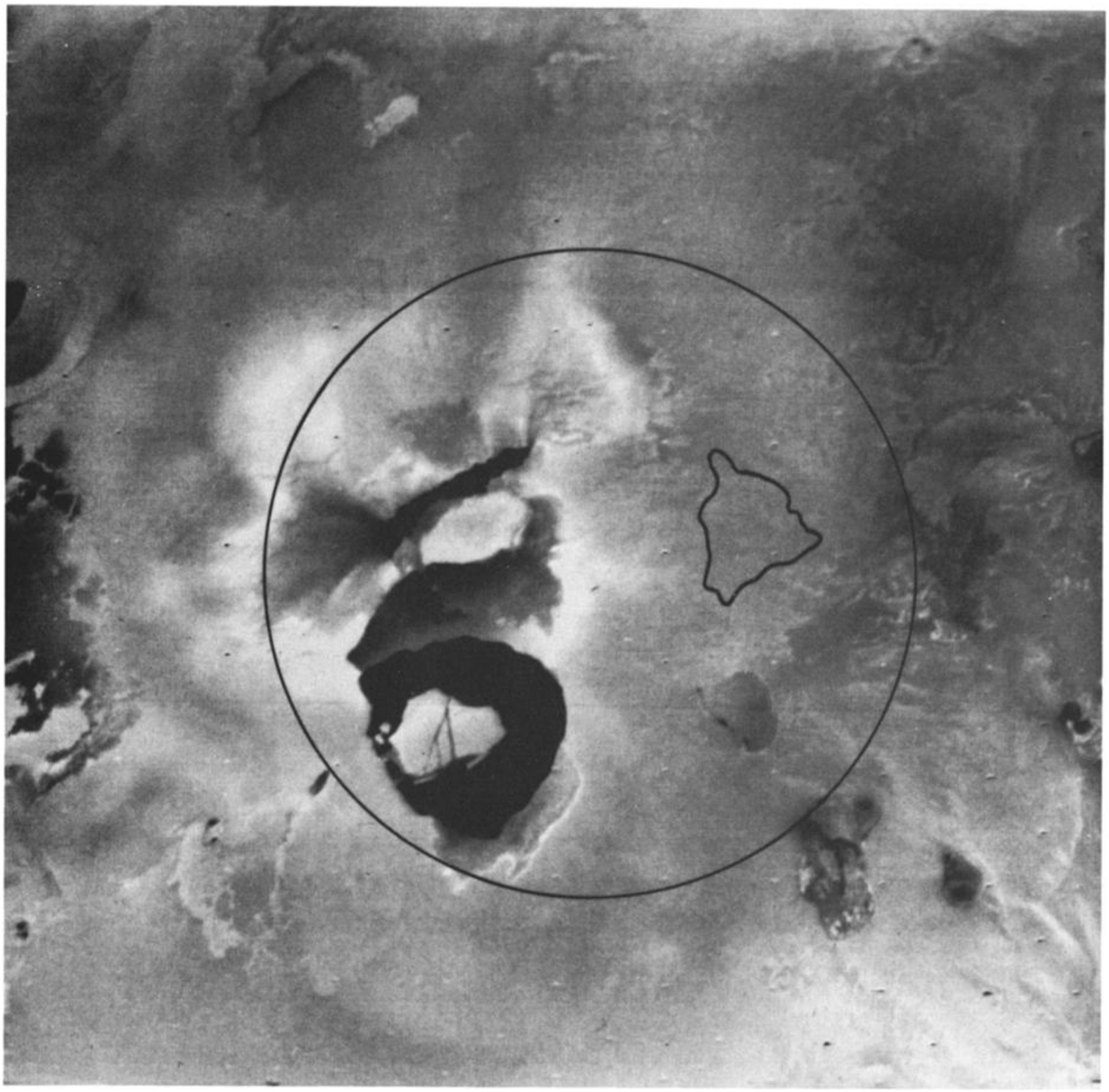

Fig. 4. The hot spot Loki compared with the field of view of the Voyager IRIS instrument. The dark horseshoe-shaped feature is 170 km across. For comparison, an outline of the island of Hawaii is shown at the same scale.

a region of sustained eruption of basalts we will briefly examine what temperatures would be detected from an active basaltic lava lake. The behavior of basaltic lava lakes is complex (see, for example, *Swanson et al.* [1979] and *Duffield* [1972]), and surface temperatures are correspondingly difficult to predict. In all but the most vigorously overturning lakes, the lake crusts over rapidly. The crust may be continually moving, spreading away from linear upwelling zones and being digested elsewhere, but even so, a crust forms within seconds so that most of lake appears crusted over [*Duffield*, 1972]. Occasionally, the entire lake will overturn, exposing uncrusted lava, but again this state lasts for only a few seconds. Thus, in general, on an active lake, only a minute fraction of the surface is at temperatures approaching liquid lava. Figure 5 gives an indication of how rapidly the surface temperatures would fall on newly exposed lava. The rate at which the lava surface cools depends on the thermal conductivity of the crust that forms, which in turn depends on its porosity. Cooling curves were calculated numerically assuming an initial lava temperature of 1450°K and a surface emissivity of 1 and assuming that the temperature at the base of the crust is maintained at 1450°K by offsetting conductive heat losses by freezing of new lava to the base of the rapidly forming crust. A heat of fusion of 380 J g^{-1} and a heat capacity of 1.2 J g^{-1} $°K^{-1}$ were assumed, and thermal conductivities for different porosities were taken from *Robertson and Peck* [1974]. The curves, particularly the 50% porosity curve, are consistent with numerous observations that newly exposed lava on a lake rapidly crusts over and becomes black within seconds and with the observation that the crust that forms is highly vesicular (W. A. Duffield, personal communication, 1985; J. G. Moore, personal communication, 1985.)

Somewhat paradoxically the high-temperature sources on Io are more difficult to explain by basaltic lava lakes than the low-temperature sources. The temperature of the surface of an active lava lake depends on the turnover rate. From the accounts of *Duffield* [1972] and *Swanson et al.* [1979] the residence time of crust on the lava lake at Maunu Ulu during 1970 and 1971 was typically of the order of hundreds of seconds. From the curves in Figure 4, temperatures should range

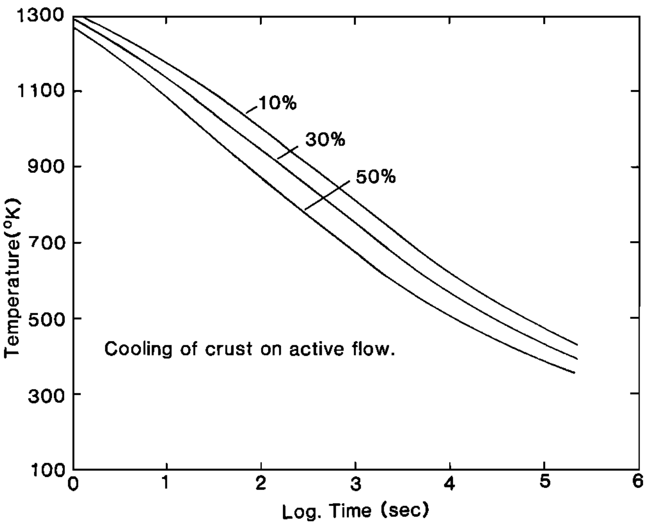

Fig. 5. Calculated temperatures of the surface of the crust on an active basaltic flow as a function of time since exposure. Curves are for different vesicularities as indicated.

from between 700°–800°K up to 1100°–1300°K, with a strong bias toward the bottom end of the scale. To get stable 600°K temperatures would require a relatively slow turnover of the lake on scale of hours, which is rarely observed on the earth. Moreover, even such a slowly overturning lake would still have a significant high-temperature component (>600°K) which was not observed by the Voyager IRIS. We can conclude therefore that the warm dark areas are unlikely to be active basaltic lava lakes. Some of the numerous low-temperature (150°–400°K) sources could, however, be inactive basaltic lava lakes. Following an eruption, the surface of a lava lake would cool as in Figure 5, and temperatures would remain in excess of 180°K for years. However, as explained below, if the anomalies are the result of silicate volcanism, prolonged eruption of flows is a more probable cause of the thermal anomalies than inactive lakes.

Terrestrial calderas are rarely, if ever, filled with molten lava. The floors of the summit calderas of the most active volcanoes on earth, Kilauea and Mauna Loa, are covered mainly by flows. Kilauea has periodically contained an active lava lake within Halemaumau, but the pit constitutes less than one twentieth of the area of the caldera floor. The floor of the Mauna Loa caldera is entirely surfaced by flows. It appears reasonable therefore to consider the possibility that the floors of the Ionian calderas are dark and warm, not as a result of the presence of lava lakes as has been widely assumed but because they are covered by relatively recent and still cooling lava flows. A suggestion that this may be so is that many of the dark areas within calderas, for example, that within Creidne Patera, have lobate outlines. In addition, different parts of the floor have different albedo, as would be expected from basalt flows, which would become brighter with time as a result of fumarolic activity and the rain of debris from the plumes. It is more difficult to explain the differences in albedo if the darkest albedo within the caldera is due to the presence of a sulfur lava lake, since once the lake stops overturning is should revert to S_8. The albedo should be either that of liquid sulfur or that of solid sulfur, but what is observed is a continuum of albedos [*McEwan et al.*, 1985]. On the other hand, no flows are visible on the floor of Loki, the most energetic of the hot spots, so hot spots may not all result from the same cause. We will explore what surface temperatures are to be expected as a result of sustained silicate volcanism and determine whether the predicted temperatures are in accord with what is observed. Unfortunately, we have no published data on the integrated thermal emission of terrestrial areas containing both active and recently active flows, so that the thermal emission will be modeled numerically. The intent is merely to determine the form of the infrared spectrum. Exact duplication of the observed spectra is not expected because surface temperatures are dependent on a wide variety of unknown factors such as the timing of eruptions, the volume of each eruption, the eruption rate, and the thickness of the flows.

A basic property of silicate eruptions that is fundamental for understanding how a silicate eruption might be detected on Io, is that most of the eruptive energy is initially buried and subsequently lost at low temperatures over several years. The excess energy E per unit volume of molten lava over solidified lava at the ambient temperature is

$$E = \rho[H_f + C_P(T_E - T_A)] \tag{1}$$

where ρ is the density of the lava (2.8 g cm^{-3}), H_f is the latent heat of fusion (380 J g^{-1}), C_P is the heat capacity (1.2 J g^{-1} °K^{-1}), and T_E and T_A are the eruption temperatures (1450°K) and the ambient temperature (100°K), respectively. If we assume that the flow stops when the viscosity falls from its initial 10^2–10^3 P near the vent to around 10^{10} P, then it would be about 60% crystalline and at a temperature of about 1340°K [*Shaw*, 1969]. (The relation between viscosity, crystallinity, and temperature is poorly understood, depending on a variety of factors such as composition, gas content, and applied shear stress, so the figures are approximate.) When the flow stops, it will have lost only 14% of its initial eruptive energy. The remaining 86% would be buried in the inactive flow to be slowly lost by low-temperature emission over an extended period of time. Furthermore, because of rapid crust formation, only a small fraction of the initial 14% lost is emitted at temperatures in excess of 1000°K.

Eruption of basaltic lava is a complex process that defies precise simulation. The sequence of events during an eruption depends on wide variety of factors such as the geometry of the vent, the chemistry of the lava, the eruption temperature, the slope of the ground, the degree of topographic confinement, and many other factors. Eruptions at Kilauea generally go through the following stages [*Swanson et al.*, 1979; *Holcomb*, 1980]. Initially, the hot, fluid basalt moves away from the vent as a broad sheet. As soon as lava is exposed, it cools rapidly to form a dark skin or wrinkled crust which founders and reforms as the lava moves. If the lava supply is sustained, the flow begins to channelize as parts stagnate and cease to move. Ultimately, flow becomes restricted to a small number of channels, commonly only one, that carry the lava from the vent to the active front. Once the channels form, they become progressively more crusted over and may finally become completely covered and form lava tubes. As the eruption continues, the active front moves farther from the vent, the lava being fed to the active front through channels and lava tubes which deliver lava to the front at a temperature only slightly below that at the vent. Forward movement of the active front may be a process of budding in which skin encased lobes of lava successively form, inflate, and burst to form new lobes, or it may be a rolling action in which clinkery blocks of partly glowing rock cascade down the active front and are ridden over by the advancing flow. Between the active part of the flow and the vent is lava that has stopped flowing and which is slowly cooling.

To model the process just described in order to determine the thermal signature of a basaltic eruption on Io, the lava flow was divided into three components; a lava channel, the active flow, and the inactive flow. The eruption rate, the width

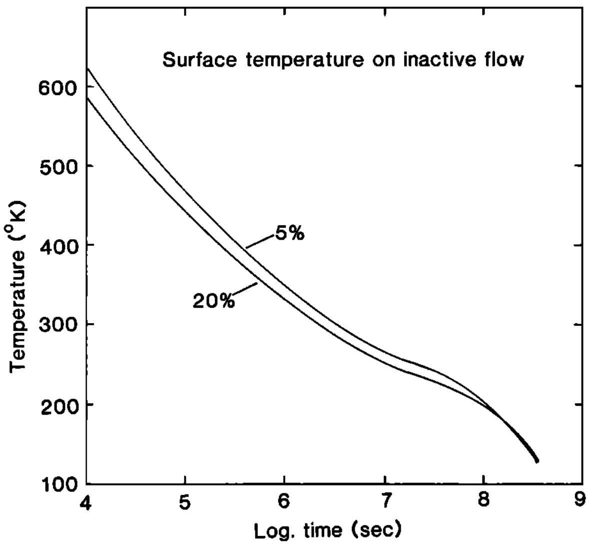

Fig. 6. Temperature of the surface of a 10-m-thick inactive basalt flow as a function of time. Curves assume that the flow is 60% crystallized when it stops flowing. The turn down in the curves after about 1 year (approximately 3×10^7 s) results because the flow at that time has completely crystallized. The curves are for different porosities as indicated.

and thickness of the lava flow, and the width of the lava channel all had to be specified. As the lava flow grows in length at a rate determined by the assumed eruption rate and the width and thickness of the lava flow, the lava channel extends itself. Temperatures were assigned to the channel by analogy with Mauna Loa. Measurements made during the 1984 eruption showed that the initial lava channel temperature was 1375°K near the vent, but because of crusting over, by the time lava within the channel had travelled 3 km, 10% of the surface was radiating at temperatures less than 775°K, and by the time the lava had traveled 20 km, 70% of the channel was radiating at temperatures less than 425°K (D. C. Pieri, personal communication, 1985). On the basis of these observations the following temperatures were assigned to the lava channel: 20% at 1375°K, 30% at 775°K, and 50% at 425°K. The length of the lava channel was also limited to 100 km on the supposition that by the time a channel reached this length it would have roofed over or reestablished itself along a new course. Heat losses from the channel substantially in excess of those calculated under these assumptions are unlikely; otherwise the viscosity of the lava would significantly increase, thereby impeding the flow and preventing the channel from conducting the lava from the vent to the active front.

To calculate temperatures on the surface of the active flow, it was assumed (1) that the active flow is continually turning over exposing new lava at the surface which rapidly cools to form a thin skin or crust, and (2) that flow stops when so much heat is lost that the viscosity of the lava rises to 10^{10} P at which time the flow is 60% crystaline, as discussed above. The fraction of the active flow at different temperatures and hence the energy loss per unit area of the flow were calculated from the cooling curves in Figure 5 and an assumed residence time of crust on the active flow. Typical residence times chosen were 200, 400, and 800 s. The thermal emission from the active flow is, however, almost independent of the residence time chosen. If the flow turns over rapidly, it cools quickly and so soon becomes inactive. Conversely, a slowly overturning flow cools slowly and so remains active over a large area. The time t that takes for the lava to freeze once it is delivered to the active front is determined by balancing the rate of loss of heat against that required to freeze 60% of the flow, taking into account also the heat lost by the lava as it travels down the lava channel to the active front:

$$t = \{d\rho[0.6H_f + C_P(T_E - T_{60})] - E_c\}/R \qquad (2)$$

where d is the flow thickness, T_{60} is the temperature at which the flow is 60% solid, E_c is the energy lost flowing through the channel normalized to the area of the active flow, and R is the rate of loss of heat per unit area of the active front. Once the time to freeze has been determined, the area of the inactive flow follows directly from the eruption rate and the flow thickness. The fractional area at different temperatures has already been established from the cooling curves so that the emission at different wavelengths can be calculated.

The rate of cooling of the partly solidified lava in the inactive flow was determined numerically as follows. All the cooling was assumed to take place from the surface which is taken to have an emissivity of 1 and to be exposed to the diurnally averaged insolation for Io (the insolation terms becomes significant only after exposure for at least a year). As the flow cools, the solid crust grows as conductive losses are offset by freezing at the base of the crust. In all the calculations a flow thickness of 10 m was assumed. This is representative of basalt flows which can range in thickness from 1 m to several tens of meters. When the entire flow freezes (almost a year after it stops moving for a 10-m-thick flow), cooling proceeds without the complications of freezing and crustal growth. A thermal conductivity of 0.015 J cm^{-1} °K^{-1} s^{-1} was used, the value for basalt with 5% porosity [*Robertson and Peck*, 1974]. The cooling curve so derived is shown in Figure 6. A lower porosity was assumed for the inactive bulk flow, as compared with the newly formed crust on the active flow, because extensive outgassing normally takes place during emplacement of a lava flow and during its subsequent cooling and because a thermal conductivity appropriate for a low porosity is consistent with the observed long-term cooling of lava lakes [*Wright et al.*, 1976]. As the eruption proceeds, larger areas are covered with basalt which has a surface temperature dependent on how long that part of the flow has been inactive. To determine the thermal emission at a particular time after the eruption started, the inactive flow was divided into segments according to how long they had been emplaced and a temperature assigned to that segment according to Figure 6. Areas and temperatures from a typical run are shown in Table 2.

Even with high eruption rates over long periods of time, the area covered with newly formed basalt is still only a small fraction of the field of view of the Voyager IRIS instrument. For example, if Loki erupted at a rate of 2000 m^3 s^{-1} for 10 years to form 10-m-thick flows, only 10% of the IRIS view of the Loki area would be covered with new flows in the 10 years, even assuming no overlap. The rest of the field would be at the background temperature. This was assumed to be the same as low-temperature component in the *Pearl and Sinton* [1982] fits to the IRIS data. In their fits, 0.856 of the field of view of Loki was at 116°K; 0.899 of the field of view of Pele was at 114°K. A temperature of 115°K was accordingly assigned to the background.

If the thermal anomalies are caused by eruptions of silicates, the eruption rates are high compared with terrestrial rates. From the *Pearl and Sinton* [1982] fits to the IRIS data (5.3 $\times 10^4$ km^2 at 248°K; 1.4 $\times 10^3$ km^2 at 458°K), 1.5 $\times 10^{13}$ W is being emitted by Loki. The equivalent figure for Pele (2.0 $\times 10^4$ km^2 at 175°K; 110 km^2 at 654°K) is 2.2 $\times 10^{12}$ W. If energy was being transported to the surface at these rates as silicate magma, the eruption rates can be calculated from equation (1), taking the ambient temperature as 100°K. The rate for Loki is 2500 m^3 s^{-1}; that for Pele is 350 m^3 s^{-1}. But as just discussed, energy brought to the surface as lava is not

TABLE 2. Typical Result of Simulation of Surface Temperatures During an Eruption

Component	Area, km^2	Temperature, °K	Power, $W\ m^{-2}$
Channel	2.00	1370	4.67×10^{11}
	3.00	775	
	5.00	425	
Active front	0.15	1210	1.61×10^{12}
	0.46	1140	
	1.37	1060	
	4.11	974	
	12.4	885	
	37.1	795	
Inactive flow	46.4	421	6.63×10^{12}
	92.9	387	
	186	356	
	371	328	
	743	302	
	1,450	278	
	2,970	257	
	5,940	236	
	11,900	216	
	23,800	197	
Rate of energy release from affected area			8.71×10^{12}.

Temperatures and areas are listed for the three components of the flow, the channel, the active front, and the inactive flow, at a time 5 years after the eruption started. The power being emitted by each flow component is also listed. Conditions are eruption rate of 3000 $m^3\ s^{-1}$, flow thickness of 10 m, turnover time on active front of 400 s, and porosity of 0.5 at surface of active front.

released instantaneously but is lost over several years. Unless eruptive activity has been sustained for several years so that equilibrium has been achieved between energy supply and energy loss, the current eruption rate should be higher than the rate implied by the energy being emitted. We should therefore expect eruption rates at Loki to be in excess of 2500 $m^3\ s^{-1}$. For comparison, the eruption rate at Maunu Ulu over a 4-year period averaged 2.9 $m^3\ s^{-1}$, with peak rates reaching 400 $m^3\ s^{-1}$ [*Swanson et al.*, 1979]. Eruption rates of terrestrial flood basalts can, however, be as high as 3000 $m^3\ s^{-1}$ [*Whitford-Stark*, 1982], but such rates are sustained for only short periods of time. Eruption rates for some lunar mare basalts may have been as high as $3 \times 10^5\ m^3\ s^{-1}$ [*Head and Wilson*, 1981].

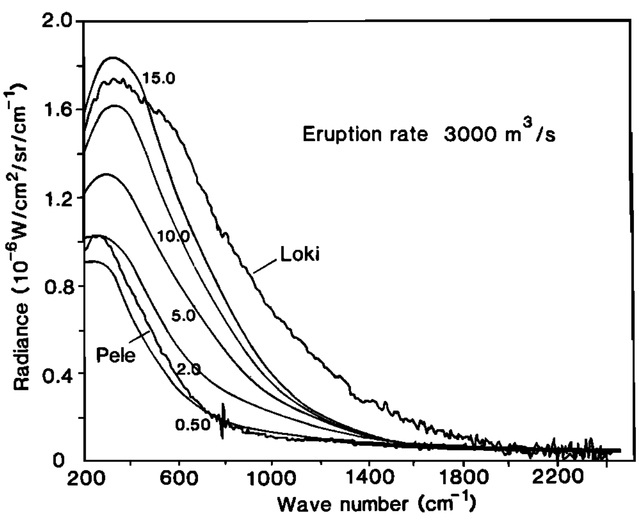

Fig. 7. Calculated radiance from an area of active basaltic volcanism campared with what was observed from Loki and Pele by the Voyager IRIS experiment. Figures by the calculated curves indicate the number of years that the eruption rate of 3000 $m^3\ s^{-1}$ was sustained.

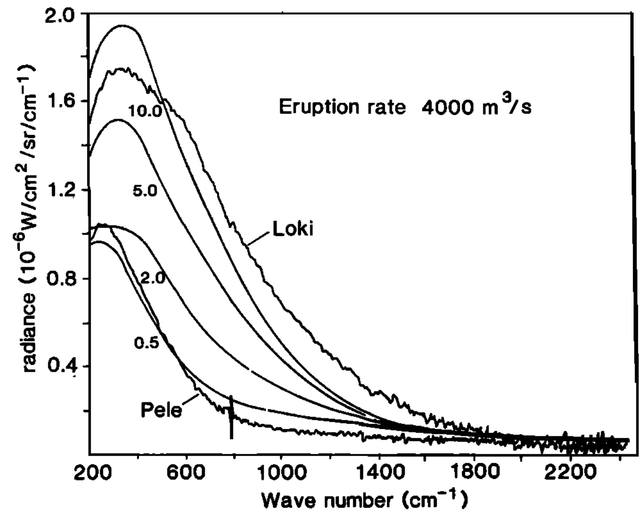

Fig. 8. As in Figure 7 except for an eruption rate of 4000 $m^3\ s^{-1}$. The observed radiance at the warm end of the spectrum is significantly higher than what was calculated, which suggests that the eruption rate could not have been as high as 4000 $m^3\ s^{-1}$ when the IRIS observations were made.

Figures 7 and 8 compare some results of numerically modeling basaltic eruptions on Io, as described above, with what was observed by the Voyager IRIS experiment. The general form of the calculated curves is similar to the observed curves but differences are apparent. Exact duplication should not be expected. A large number of simplifying assumptions were made in the modeling. Variations in the eruption rate, superposition of flows on flows, changes in flow thickness, transition from aa to pahoehoe, outgassing, and other commmon characteristics of terrestrial eruptions were not modeled. Nevertheless, the agreement between the curve for Pele and that expected from an eruption of 3000 $m^3\ s^{-1}$ sustained for 6 months is quite good. The agreement of the calculated curves with what is observed from Loki is not as good. Sustained high eruption rates are required to explain the part of the curve between 600 and 1800 cm^{-1}, but such high rates should be more evident in the signal from the active parts of the flow ($>2000\ cm^{-1}$) and result in accumulation of a greater area of relatively cold (<150°K) flows than is observed. In effect, a higher fraction of the Loki eruption site than predicted is at temperatures in the 200°–500°K range as compared to areas with 100°–200°K temperatures and areas with temperatures in excess of 500°K. The rather modest discrepancy may be explained by the presence of sulfur melts, although this has not been assessed quantitatively because of lack of detailed information on surface temperatures on cooling sulfur flows or lakes. But there are also several plausible explanations consistent with basaltic eruptions. The relatively high signal in the 200–600 cm^{-1} range, which comes from old, cold flows, is probably an artifact of the model because no account was taken of overlap of new flows onto old. This must occur so that the fraction of ground at low temperatures is almost certainly overestimated in the model. A possible explanation of the relatively low signal in the 1800–2400 cm^{-1} range compared with 600–1600 cm^{-1} range is that the eruption rate at the time of the Voyager encounter was somewhat less than the 10-year average. However, another, and possibly more plausible, explanation is that the 600–1600 cm^{-1} part of the curve is enhanced over the predicted curve because hot gases continue to be expelled from the flow after it stops moving, thereby causing local hot spots, or fumaroles, on an otherwise cool surface. We have abundant evidence from Io that volatiles, particularly sulfur compounds, condense around local vents.

Many calderas, flows, and hills are surrounded by haloes or auras which appear to be areas where locally released volatiles have condensed (S. M. Baloga et al., unpublished manuscript, 1986). Release of hot gases and their condensation to form fumarolic deposits would result in an increase in the fraction of ground at temperatures in the 200°–500°K range, as compared with those predicted by the simply cooling model. A suggestion that this is occurring in Loki is the presence of numerous bright spots throughout the dark area. These have been previously interpreted as islands within a lake of molten sulfur cover, but sites of fumarolic activity within an area of basaltic eruptions appear equally plausible.

Thus while the IRIS measurements differ in detail from those predicted for basaltic eruptions, the general form of the temperature distributions observed is consistent with basalts, and the deviations in detail are explicable in terms of familiar processes that occur in terrestrial eruptions. Despite an attempt to model the cooling of a sulfur lava lake by *Sinton* [1980], the temperature distribution that would result from sulfur flows and lakes is largely unknown because of the lack of direct observations of the rate of convective overturn or the details of flow emplacement. *Lunine and Stevenson* [1986] have, however, modeled the thermal emission from sulfur lava lakes heated from below by hot silicates, and their predictions are in good agreement with the IRIS measurements. The thermal measurements are therefore equivocal, being explicable both in terms of sulfur and silicate volcanism or a combination of the two.

IONIAN VOLCANISM

We have seen that if the crust of Io is more than 20 km thick, as appears probable, a significant amount of its tidal energy must be dissipated below the lithosphere and be transported toward the surface as silicate magma. We have also seen that the thermal anomalies are consistent with basaltic eruptions, so it appears reasonable that some silicate magma could reach the surface. *Smith et al.* [1979*b*] raised two arguments against any silicate volcanism. First, they claimed that silicates are not buoyant enough to rise through the sulfur-rich materials at the surface. Second, they suggested that if silicates were erupted onto the surface, they would irreversibly remove all the sulfur from the surface over geologic time. Since this has not happened, they claim that silicate volcanism does not occur to any significant degree. The buoyancy argument against silicate volcanism raised by *Smith et al.* [1979*b*] is somewhat circular. If indeed Io had a sulfur crust several kilometers thick, buoyancy would prevent silicates reaching the surface. But a sulfur-rich crust presupposes that silicates do not reach the surface, so the conclusion is not unexpected. If the surface is dominantly silicate, as is implied by its strength [*Clow and Carr*, 1980], then silicates do reach the surface. Magma is believed to rise to the surface largely as a result of differences in pressure exerted at the magma source by the rock column and the magma column (see, for example, *Holmes* [1945, p. 478] and *Eaton and Murata* [1960]). If we assume an all-basalt lithosphere, a surface density of 3.0 g cm^{-3}, and a coefficient of volume expansion of 9.6×10^{-5}, the density at the base of the lithosphere is 2.89 and its average density (ρ_a) is 2.95 g cm^{-3}. The change in volume on melting (ΔV_m) can be estimated from the Clausius Clapeyron equation

$$\Delta T/\Delta P = T(\Delta V_m)/\Delta H_m \tag{3}$$

Taking $\Delta T/\Delta P = 5°\text{K kbar}^{-1}$, $T = 1420°\text{K}$, and ΔH_m, the enthalpy of melting, $= 90$ cal g^{-1} [*Yoder*, 1976], the change in volume on melting is 0.0132 cm^3 g^{-1}, and the density of the lava (ρ_m) is 2.78 g cm^{-3}. Balancing the lava column and the rock column, the height limit for volcano growth (h_v) is $L(\rho_a - \rho_m)/\rho_m$, where L is the lithosphere thickness. By this reasoning a volcano on Io can grow no higher than 1.8 km if the lithosphere is basaltic and 30 km thick and no higher than 3.6 km on a 60-km-thick lithosphere. If the lithosphere includes a dense fraction such as peridotite, volcanoes could grow higher, but Io is likely to be highly differentiated and have a lithosphere consisting entirely of a low melting temperature fraction such as basalt. The low height limit is consistent with the general absence of large shield volcanoes such as occur on Venus, Earth, and Mars, despite high rates of volcanism. Occasional shield volcanoes about 2 km high do occur, however [*Moore et al.*, 1984].

Even if the near-surface materials contained a large fraction of a low-density component such as sulfur, the pressure differential between the lava and rock columns could still pump the magma to the surface. For example a 5-km-thick sulfur layer over 25 km of basalt almost balances a 30-km column of basalt lava. However, in this case the lava would tend to be intruded under the sulfur and not reach the surface. Whether extrusion or intrusion occurs depends on stress conditions near the surface, but in the ideal case of no horizontal stresses and negligible strength of the magma conduit walls, extrusion will result only when the lava is less dense than the surface rocks. Addition of 22% sulfur with a density of 2.0 g cm^{-1} to solidified basalt will result in a density identical to basaltic lava. Thus the presence of more than about 20% sulfur will tend to suppress extrusion in favor of intrusion. Given that some silicate volcanism occurs, this could be viewed as a rough upper limit on the amount of sulfur that could be present in the upper few hundred meters of the surface. For comparison, *Clow and Carr* [1980] concluded that the upper 2 km could not contain more than several percent sulfur; otherwise the 2-km-deep calderas would collapse.

The argument that iron-rich silicate magmas cannot be extruded onto the surface because they would react with sulfur and ultimately remove the sulfur from the surface cannot be disproved, but if the process occurs at all, it is likely to be inefficient. Scavenging requires that sulfur be transported to the base of the lithosphere in sufficient quantities that the solubility limit of sulfur in silicate magma, which is about 1% [*Haughton et al.*, 1974], is exceeded. The excess sulfur would then form a dense, iron-rich sulfide melt. This would accumulate at the base of the zone in which silicates melt and be removed from further participation in surface processes. Sulfur could be transported to the base of the lithosphere in two ways. First, the sulfur could chemically react with lava near the surface to form sulfides. Second, the sulfur near the surface could simply be mixed with silicates. In both cases, subsequent burial would slowly carry the sulfur downward through the lithosphere (at a rate of a few tenths of a centimeter per year if Figure 3 is correct). However, chemical reaction between silicates and sulfur at the surface is likely to be trivial. Intrusion of silicate lava into a cold, sulfur-rich host would cause immediate quenching of a thin skin, which would form a protective rind between the lava and the sulfur, thereby preventing further chemical reaction. The situation is analogous to the reaction beween basalt and seawater during undersea eruptions or when lava flows into the sea. The protective rind prevents seawater from entering the basalt, and the lavas remain undersaturated in water; the little water that is present in the basalts after solidification was probably there before eruption [*Moore*, 1965]. If the sulfur is simply mixed with basalt and then buried, it will be hindered from reaching the base of the lithosphere by melting and vaporization. At shallow depths the sulfur will melt and tend to migrate upward through the overlying rocks. At deeper, hotter parts of the lithosphere, the sulfur will tend to be vaporized and driven upward. Both melting and vaporization would be accelerated by intrusion of hot silicates from below. Whether significant amounts of sulfur could reach the base of the lithosphere and be removed from further circulation would depend on how the rates of upward diffusion compare with burial rates, which we have no way of estimating. However, the process is so uncertain that it cannot be viewed as a strong argument against silicate volcanism.

We have been mostly concerned above with the case for silicate volcanism, but other materials are clearly present and other forms of volcanism must occur. Strength arguments suggest that silicates constitute the bulk of the near-surface materials and hence the bulk of the erupted materials [*Clow and Carr*, 1980], but several percent of other materials may be present. SO_2 has been unambiguously identified, and other sulfur compounds such as elemental S and S_2O [*Hapke*, 1979] may also be present. In addition, sodium and potassium compounds must be present to account for the sodium and potassium clouds in Io's vicinity [*Trafton et al.*, 1974; *Trafton*, 1975]. Thus Io's surface is probably complex with silicates interbedded with a variety of other materials. The volcanism is likely to be correspondingly complex. Eruption of silicate magma into the near-surface materials, where significant fractions of volatile component may be present, will result in mobilization of these components. This may result in explosive eruptions, such as plumes, or quiet eruptions of lava onto the surface.

Whether sulfur lava lakes or sulfur flows cover a significant fraction of the surface is unclear. Eruption of liquid sulfur onto Io's surface has recently been questioned by *Young* [1984], who shows that the supposed color match between various orange and red allotropes of sulfur and different parts of Io's surface (see, for example, *Morrison* [1982, p. 937]) is completely spurious. Io is yellowish-green, not orange-red. Furthermore, the only allotrope which is stable at Io's surface temperature is S_8. However, while all other allotropes will revert to S_8, the process may be so slow that allotropes such as S_3 and S_4 could survive on the surface for days, possibly years [*Gradie and Moses*, 1983]. Young also points out that Io is much brighter in the blue and UV than S_8. Also troublesome is the lack of observation of a temperature dependent shift in the absorption edge between 0.4 and 0.5 μm, which is characteristic of sulfur [*Gradie et al.*, 1982; *Gradie and Veverka*, 1984; *Veverka et al.*, 1982]. The discrepancy between the reflectance spectra of Io and S_8 is so large that Young regards sulfur as an unlikely cause of the surface colors. Nevertheless, while the presence of allotropes other than S_8 may be doubtful, many of the arguments for sulfur remain valid. Io must be the source for the sulfur in the torus, SO_2 has been positively identified and is widespread, and while there are differences in reflectivity between Io and sulfur, the comparisons are based on pure sulfur, and contaminants within the sulfur could significantly change its spectral reflectance [*Gradie and Moses*, 1983; *Soderblom et al.*, 1980]. In addition, *Pieri et al.* [1984] find color changes along flows from Ra Patera that are suggestive of transition of sulfur through different allotropes as it cools. Moreover, if the surface contains a few percent sulfur, local sulfur flows and possibly lakes should be expected.

What we know of Io is thus consistent with the following. Its volcanism and hence its surface materials are dominantly silicic. Several percent of volatile materials such as sulfur, but also including sodium- and potassium-rich materials, may also be present. The volatile materials at the surface are continually vaporized and melted as a result of the high rates of silicate volcanism. As the erupted products become buried, as a result of ongoing eruption of new material onto the surface, their temperatures rise. Volatile components are melted and volatilized at relatively shallow depths and transported upward, possibly to be erupted again onto the surface. As a consequence of their mobility, the volatile components become progressively depleted with depth. Whether they reach the depths where silicates melt will depend on their rate of diffusion and convection through the overlying lithosphere. Most of the sulfur compounds are likely to be depleted at relatively shallow depths because of their high volatility and so are recycled on a time scale that is short compared with the silicates.

If all the silicate magma generated were erupted rather than intruded, the surface would be covered at a rate of a few tenths of a centimeter per year. Although unrealistic, such a rate is consistent with the burial rates of craters [*Johnson et al.*, 1979; *Johnson and Soderblom*, 1982] and is substantially higher than the resurfacing rates of the currently active plumes, as must be the case if the visible flows and the terrain are dominantly silicate. The resurfacing rates imply that the entire lithosphere is overturned every few tens of millions of years, or about 100 times in the history of the planet.

McEwan and Soderblom [1983] showed that in the 4 months between Voyager encounters, the two calderas Aten and Surt had changed and become very dark, and major changes had taken place at Loki. However, few, if any, changes in the flow pattern between calderas were observed between encounters. Flows around Ra Patera typically have a surface area of 2000 km^2 [*Pieri et al.*, 1984]. If they are about 10 m thick, as might be expected for a basalt, then formation of one flow would be equivalent to resurfacing the entire planet to a depth of 0.05 cm. Thus if all the magma estimated to be produced (Figure 3) was erupted as silicate flows, some changes in the flow pattern should have been observed. Because of their low viscosity, terrestrial sulfur flows are typically only a few centimeters thick [*Watanabe*, 1940; *Skinner*, 1970; *Naranjo*, 1985]. If all the energy losses were accomplished by sulfur flows similar in thickness to those on earth and comparable in areal extent to the Ra Patera flows, then hundreds should have formed between the two Voyager encounters. The lack of change supports the supposition that most of the volcanism is of silicates but also suggests that most of the resurfacing is accomplished by formation and filling of calderas rather than resurfacing of the intercaldera areas.

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