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A post-Galileo view of Io's interior

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Abstract

We present a self-consistent model for the interior of Io, taking the recent Galileo data into account. In this model, Io has a completely molten core, substantially molten mantle, and a very cold lithosphere. Heat from magmatic activity can mobilize volatile compounds such as SO_2 in the lithosphere, and the movement of such cryogenic fluids may be important in the formation of surface features including sapping scarps and paterae.
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1. Introduction

Io is the most geologically active body in the Solar System. Not surprisingly, the speculation on the processes underlying this activity is nearly as dynamic as Io itself. The primary goal of this paper is to summarize recent results related to Galileo observations and to present a preliminary synthesis of these new results. What is presented here is far from a consensus view of all Io researchers, but it does provide a self-consistent model that fits the available observations. In fact, the model is highly controversial in that it challenges some commonly held concepts, especially about magma migration, mantle dynamics, and geomorphologic interpretation. We begin with a brief synopsis of the evolution of thinking about the interior of Io, before and during the Galileo spacecraft's seven-year tour of the jovian system.

1.1. Pre-Galileo ideas

Io was first discovered by Galileo Galilei on January 7, 1610 using one of the earliest Earth-based telescopes (Galilei, 1610). Since then, improved telescopic observations have shown that Io is a world unlike any other in our Solar System, appearing anomalous in both the visible and infrared (e.g., Wamsteker et al., 1974; Morrison, 1977). The two Voyager spacecraft provided the first detailed look at Io, revealing active volcanism in the form of infrared hot spots and gaseous plumes reaching hundreds of kilometers into space (e.g., Morabito et al., 1979; Hanel et al., 1979). Active volcanism on Io was predicted just months before the Voyager flybys, based on calculations of the tidal heating within Io as it travels in an orbit resonant with Europa (Peale et al., 1979). Since the Voyager flybys, Io has been extensively monitored by Earth-based telescopes, including the Hubble Space Telescope (e.g., Spencer et al., 1997). These diverse data sets provided fuel for interesting discussions about the interior processes of Io.

1.1.1. *Sulfur versus silicate volcanism*

While Voyager provided irrefutable proof that Io was volcanically active, the composition of the lavas was not established with certainty. The colors of the lavas seen in the Voyager images suggested sulfur (Sagan, 1979), and the temperature estimates from the Infrared Radiometer Interferometer and Spectrometer (IRIS) instrument could all be explained by sulfur (Sinton, 1982). However, it was also known that the colors could be due to thin surface coatings and that the IRIS data could also be consistent with cooling silicate (presumably basaltic) lavas (e.g., Carr et al., 1979; Carr, 1986). The consensus shortly after the Voyager flybys was that the volcanism seen at the surface of Io was dom-

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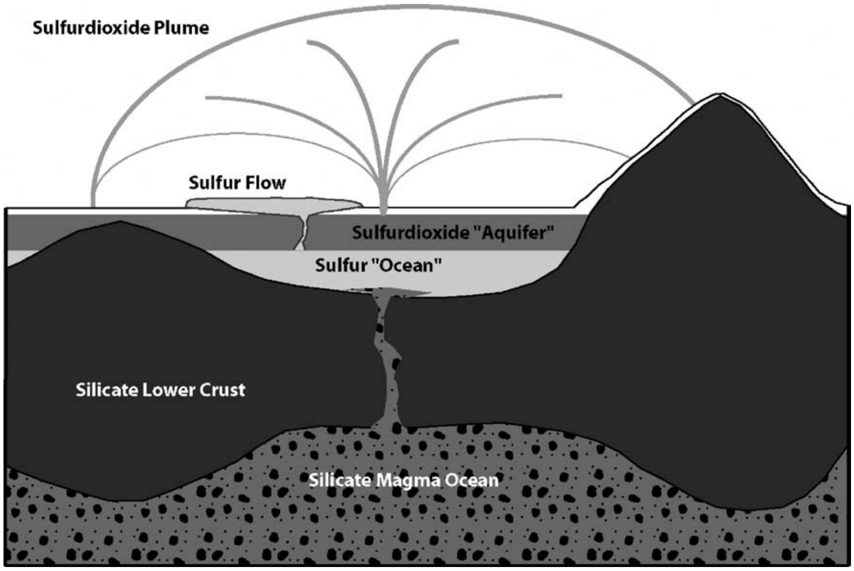

Fig. 1. A generally accepted view of the interior of Io immediately after the Voyager flybys. After Smith et al. (1979b). The intense tidal heating was thought to be sufficient to completely melt sulfur at a depth of about a kilometer and silicates at a depth of $<$ 20 kilometers. Volcanism was thought to be predominantly sulfur with SO_2-rich plumes.

inated by sulfur lavas, but that this sulfur volcanism was driven by underlying silicate magmas (Smith et al., 1979a) (Fig. 1). However, later ground-based telescopic observations of infrared emissions from Io showed that temperatures in excess of the boiling point of sulfur occurred on several occasions (e.g., Johnson et al., 1988; Veeder et al., 1994). Thus, by the time Galileo arrived in Jupiter orbit, it was generally accepted that active silicate (i.e., basaltic) and sulfur volcanism would be seen on Io's surface (e.g., Spencer and Schneider, 1996).

1.1.2. The rise and fall of the magma ocean

The nature of the source of the silicate magma was also a matter of considerable discussion. The initial thermal modeling of Peale et al. (1979) predicted that the interior of Io could be completely molten. In their model, the tidal heating was dominantly caused by tidal flexing of a thin (few kilometers thick) mechanical lithosphere. However, as this suggestion was studied in more detail, a series of problems with the magma ocean were identified. First, the towering mountains on Io (reaching $>$ 17 km in height (Schenk et al., 2001)) could not exist if the lithosphere were so thin (O'Reilly and Davies, 1981). This problem was solved by O'Reilly and Davies (1981) who showed that the resurfacing of Io is so rapid that new (cold) lithosphere is generated faster than it can be heated by conduction from below. Thus, exactly because Io is so volcanically active, a thick and very cold lithosphere is expected.

With the thicker and more rigid lithosphere, the observed tidal heating could no longer be generated in the lithosphere. New models examined deformation of the mantle as well as the lithosphere and found that the necessary heat could be generated throughout the interior of Io (e.g., Ross and Schubert, 1985; Ojakangas and Stevenson, 1986). However, the distribution of heating depended on the details of the interior structure (Segatz et al., 1988). One hope for observationally constraining the interior structure of Io came from the realization that the spatial distribution of heating would be very different depending on the depth where the heating was concentrated. Deep heating would result in maximum heat generation (and presumably volcanism) at the poles while shallower asthenospheric heating would be concentrated at the sub- and anti-jovian points (Ross et al., 1990). While the Voyager imaging of Io was less than complete, the distribution of volcanic centers on Io suggested that most of the heating was in the asthenosphere with a non-trivial contribution from deep in the mantle (Ross et al., 1990). This led to the general view that Io was a largely solid body with small degrees of partial melting concentrated in the upper mantle but extending into the lower mantle (Fig. 2). This was dubbed the "asthenospheric heating" model (Ross et al., 1990).

It is interesting to note that models of the orbital evolution of Io suggest that the current level of tidal heating is not sustainable over the age of the Solar System because Io would have to have formed within Jupiter (e.g., Schubert et al., 1986). The only way to avoid this result is if the interior structure of Jupiter is fundamentally different from what we expect. Coupled models for tidal evolution and tidal heating within Io allow for periodic episodes of high-intensity heating, separated by longer intervals of relatively little activity (Ojakangas and Stevenson, 1986).

1.2. The Galileo mission(s)

The science reaped from the Galileo data is a testament to decades of effort by hundreds of individuals. It is impossible to discuss the discoveries provided by Galileo, and their

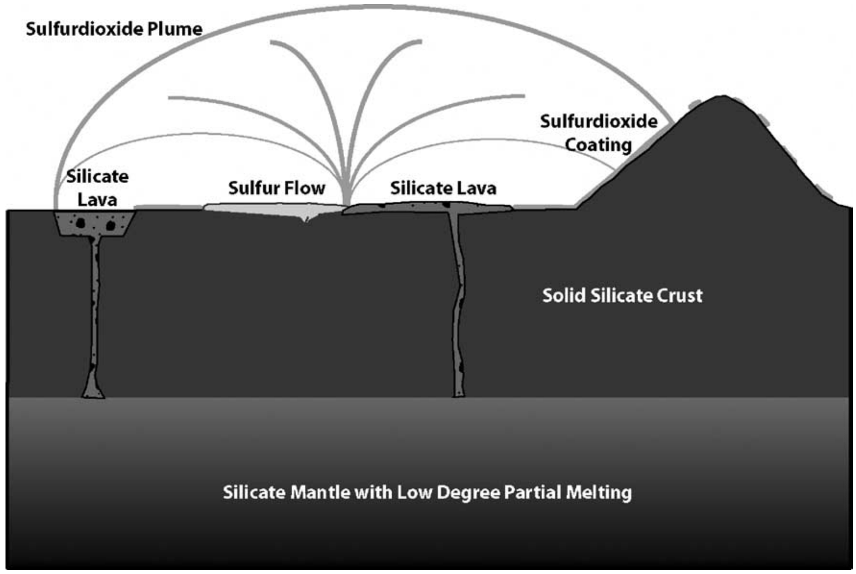

Fig. 2. A generally accepted view of the interior of Io immediately before Galileo arrived at Jupiter. After Spencer and Schneider (1996). The silicate portion of Io was presumed to be mostly solid with only low degree partial melting of the mantle, mostly in the uppermost $\sim$ 50 km. Sulfur compounds were thought to be largely confined to the uppermost crust and in a surface deposit that covered most of the silicate rocks. Volcanism was seen to be a combination of sulfur and silicate lavas. The lithosphere was generally estimated to be $\geqslant$ 30 km.

implications, without a brief summary of the history of this mission and the different types of instruments onboard the two-ton spacecraft.

1.2.1. Brief history

The Galileo Mission was conceived in the mid-1970s, even before the Voyager flybys of Jupiter. After a convoluted trip to the launch pad and through the Solar System (returning by Earth twice and flying by Venus and two asteroids), Galileo entered orbit around Jupiter in December of 1995. The prime mission lasted only 2 years, but was immediately followed by two extended missions (the Galileo Europa Mission (GEM) in 1997–1999 and the Galileo Millennium Mission (GMM) in 1999–2003) (see Turtle et al., 2004). Galileo completed 34 orbits around Jupiter before burning up in the jovian atmosphere in September 2003. Galileo's observations are sorted by orbit, which are named by both the target of the close flyby and the number of the orbit. Thus orbit "I27" came closest to Io and was the 27th orbit of Jupiter.

For the prime mission and the first part of GEM, Galileo was only able to observe Io from a great distance (McEwen et al., 1998a). During the latter part of GEM and during GMM the spacecraft was brought deeper into Jupiter's radiation belts, allowing it to acquire the first high-resolution data from Io. However, the intense radiation and the age of the spacecraft combined to cause many difficulties during these close flybys (e.g., McEwen et al., 2000; Keszthelyi et al., 2001; Turtle et al., 2004).

1.2.2. Instrumentation

The Galileo spacecraft carried a full complement of scientific instrumentation, covering most of the electromagnetic spectrum. Of particular interest to Io investigations were the gravity and magnetometer experiments, as well as a trio of remote sensing instruments: the Solid State Imager (SSI), Near Infrared Mapping Spectrometer (NIMS), and Photopolarimeter–Radiometer (PPR). SSI collected 800 $\times$ 800 pixel images between 0.4 and 1.0 µm using 8 filters. Image resolutions varied from tens of kilometers per pixel to 5 m/pixel (Turtle et al., 2004). NIMS covered between 0.7 and 5.2 µm and built up images in vertical strips. Spectral information was obtained by a combination of 17 detectors and 24 grating positions, allowing 408 samples, although by the time of the Io flybys the spectral resolution had degraded to 12 positions between 1.0 and 4.7 µm (Lopes-Gautier et al., 2000). NIMS spatial resolution varied between hundreds of kilometers per pixel to hundreds of meters per pixel (Lopes et al., 2004). PPR was capable of broadband observations, from visible wavelengths to 100 µm, and built up images by scanning across the target. At Io, much of the data was collected at 17 and 21 µm and images ranged in resolution from hundreds of kilometers per pixel to a few kilometers per pixel (Spencer et al., 2000a; Rathbun et al., 2004). While each of these instruments provided valuable information individually, the greatest insights have come from co-analysis of the different data sets.

2. Galileo discoveries at Io

We will not attempt to list all of the discoveries at Io that Galileo produced in more than seven years in orbit around Jupiter. Instead, here we only highlight the key observations that have changed our thinking about the interior of Io.

2.1. High-temperature volcanism

Perhaps the most unexpected result from Galileo was the sight of dozens of spots of incandescent lava imaged from $> 100{,}000$ km (> 1 km/pixel resolution) by the SSI camera while Io was in eclipse. In order to be visible to SSI, the lava surfaces must have been > 700 K, well above the vaporization temperature of sulfur (~ 400 K). For the observed ten-kilometer-scale areas to be above 700 K, the interior of the lava flows must have been > 1000 K, in the range of typical silicate lavas (McEwen et al., 1997, 1998a). While earlier Earth-based telescopic observations indicated that silicate lavas erupted on Io from time to time, the Galileo SSI and NIMS data showed that silicate lavas were continually erupting in many locations (e.g., Lopes-Gautier et al., 1999).

For more precise estimates of the lava temperatures (and thus the lava composition), co-analysis of NIMS and SSI data proved to be the most useful. While NIMS provided good coverage of the thermal emission spectrum from ionian hot spots, SSI provided important additional constraints on the short-wavelength emissions. It is the short-wavelength end of the spectrum that most tightly constrains the high-temperature end of model fits to the emission spectrum. Fits to the thermal emission from most of the ionian hot spots provided temperatures consistent with basaltic lavas (1200–1400 K), although higher temperatures could not be ruled out (e.g., Davies et al., 2000). The extremely energetic 1997 eruption at Pillan Patera provided the highest well-constrained temperature estimate. The emissions from this eruption are best fit with a model that has an initial lava temperature of 1870 ± 25 K. This temperature is well above the range for basaltic lavas and suggests ultramafic compositions (McEwen et al., 1998b; Davies et al., 2001).

The possibility that most of Io's silicate volcanism is actually ultramafic, rather than basaltic, in composition is supported by SSI color data. The one-to-one correlation between hot areas and dark areas on Io first seen in the Voyager data (McEwen et al., 1985) was also seen in the Galileo data (Geissler et al., 1999). Furthermore, all of the dark areas were darkest in the 968 nm filter, suggesting an absorption band very close to 1 μm. The best fit to this absorption band (and the full 6-color data from the dark areas) was extremely magnesian orthopyroxene (i.e., enstatite) (Geissler et al., 1999). Such magnesium-rich pyroxenes require ultramafic lavas, suggesting that most of the dark lavas on Io are ultramafic rather than basaltic in composition. Interestingly, the SSI color data are inconsistent with large quantities of olivine (Geissler et al., 1999), which is the most common mineral in terrestrial ultramafic lavas (e.g., Williams et al., 2000). The closest terrestrial analog is the Commondale komatiite which does include large amounts of magnesian orthopyroxene and has a dry liquidus temperature of 1884 K (Williams et al., 2000).

The presence of very high temperature lavas on Io has extremely significant implications for Io's interior. If Io has been as volcanically active as it is today (erupting ~ 550 km^3/yr of lava (Blaney et al., 1995)) for the past 4 billion years, it has produced a volume of lava equivalent to ~ 140 times the volume of Io. Even if Io were frequently much less volcanically active, we would expect the interior to be extremely differentiated by multiple episodes of partial melting. Such partial melting would inexorably lead to a low-density crust with a low melting temperature (Keszthelyi and McEwen, 1997). If the refractory portion of the mantle could be melted, the resulting melt would be predicted to be very olivine rich. The observed enstatite-rich, high-melting-temperature lavas should not exist on a highly differentiated Io. The only way such lavas could exist is to efficiently recycle the crust into the mantle. A very large degree of partial melting at the base of the crust is one mechanism to mix the crust back into the mantle (Keszthelyi et al., 1999).

Since the existence of high-temperature, enstatite-rich ultramafic lavas is critical for the model we propose (and for excluding other models), it is important to examine the limits of the observations in more detail. There are two lines of evidence that support the presence of such lavas, high temperatures and SSI color data. First, there are ways in which lava temperatures could fail to reflect the melting temperature of the lava (and thus provide no constraint on the composition of the lava). One possibility is that the lavas are superheated (i.e., are hotter than the equilibrium melting temperature at the surface pressure). If the lava were roughly basaltic in composition, superheating of nearly 500 K would be required to explain the Pillan observation. This could happen in a number of different ways. On Earth, superheating can occur when magma rises very rapidly from great depth. Since the adiabatic temperature gradient is less than that of the liquidus, the lava that is erupted is hotter than the melting temperature at the surface. However, raising a typical Kilauea basalt's melting temperature to 1870 K would require a pressure greater than that at the base of Io's mantle (~ 2–3 GPa).

It is also hypothetically possible for tidal heating to superheat liquid magma while in the conduit. However, heat added to magma would normally go toward melting more of the surrounding country rock. To overcome this buffering, the heat must be added to the liquid faster than it can be conducted away. It is relatively straightforward to estimate the conditions under which tidal superheating could occur in simple geometries, such as a dike. If we assume that the tidal heating in the dike takes the form of viscous dissipation as the two walls of the dike are moved relative to each other by tides, then the heating can be described by $Q_{\mathrm{tide}} = \eta U_{\mathrm{wall}}^2 / W_{\mathrm{dike}}$, where Q_{tide} is the tidal heating, η is viscosity, U_{wall} is the relative velocity of the two walls of the dike, and W_{dike} is the width of the dike (Turcotte and Schubert, 1982). In order for the magma to be held hotter than the surrounding rocks, this tidal heating must match the conductive heat loss from the dike, which can be given as $Q_{\text{cond}} = k(2 * \Delta T / W_{\text{dike}})$, where Q_{cond} is the conductive heat flux out of the dike, k is thermal conductivity, and ΔT is the temperature difference between the hottest magma and the wall rock. For reasonable values for k (~ 1 W/mK) and η (~ 10 Pa s) (e.g., Turcotte and Schubert, 1982), a ΔT of 500 K can be sustained only if $U_{\text{wall}} \cong 3$ m/s. Even if the tides were 100 m high and one side of the dike did not move, U_{wall} would only be 1.3×10^{-3} m/s. Therefore, we discount localized tidal heating as a plausible superheating mechanism.

One exotic superheating mechanism cannot be completely ruled out. The auroral glows around Io show that massive electrical potentials and currents surround Io. It is possible that these could reach the surface via electrically conductive gases within volcanic plumes. If the currents surrounding Io were passed through a few small (i.e., 100-m-scale) locations on Io, then there would be sufficient power to heat the ground by hundreds of degrees within a matter of minutes. Liquid basaltic lava is an appropriate electrical conductor for passing such currents. However, it must be noted that there are no observational data indicating that the auroral glows do reach the surface (Geissler et al., 2001). Furthermore, some unknown mechanism must be called upon to confine the currents to a relatively small volume of lava for this mechanism to explain the observed lava temperatures.

Therefore, we have substantial confidence in the observation that some of Io's lavas are indeed too hot to be basaltic in composition, even though we cannot absolutely rule out an exotic form of superheating.

The detection of enstatite (and the non-detection of olivine) is less definitive. There are other minerals, such as talc, that could fit the 6-color SSI data (Geissler et al., 1999). However, the lack of evidence for more than hints of hydrous minerals or water in the infrared spectrum of Io (e.g., Salama et al., 1990) makes talc an unlikely mineral to be present within the recently active lavas on Io. NIMS would have provided sufficient spectral resolution to make definitive mineral identifications, but unfortunately the key detectors in the nearest infrared were lost very early in the mission, before the close Io flybys when the instrument's spatial resolution was sufficient to isolate the recently active lavas. Thus, we have good reason to expect the presence of magnesium-rich orthopyroxenes in Io's lavas, but we are far from certain in this interpretation.

2.2. Volatile behavior

Galileo images also changed our understanding of how plumes are generated on Io. The Voyager data showed that there were at least two major types of plumes, dubbed Prometheus-type and Pele-type. Prometheus-type plumes were bright in visible light and were typically $\leqslant 100$ km tall. Pele-type plumes were distinct in ultraviolet light and were typically > 200 km tall (McEwen and Soderblom, 1983). Galileo images showed that Prometheus-type plumes are most often generated near the fronts of active lava flows. In contrast, Pele-type plumes tend to form at the vents for silicate lavas (McEwen et al., 1998a).

Ultraviolet observations of the Pele plume using the Hubble Space Telescope showed that the gases included a significant amount of elemental sulfur in addition to SO_2 (Spencer et al., 2000b). The bright red deposits from Pele-type plumes, and the other diffuse red deposits associated with the sources of silicate lavas, are also consistent with a significant amount of short-chained sulfur. S_3 and S_4 are typically produced when sulfur is vaporized at the temperatures associated with silicate volcanism. This meta-stable form of sulfur then converts to yellowish S_8 with time. Observations of the evolution of these red diffuse deposits confirmed that they lose their red color over a time scale of months, consistent with this transition (Geissler et al., 2004). Cl_2SO_2 may also play a role in the diffuse red deposits (Schmitt and Rodriguez, 2003).

In contrast to the Pele-type plumes, it appears that Prometheus-type plumes are primarily composed of SO_2. Thermal modeling suggests that the Prometheus plumes require a significant reservoir of liquid SO_2 underneath a relatively thick pile of hot lava (Kieffer et al., 2000). Small jets seen extending from the fronts of active lava flows (Fig. 3) can be explained by heating at the front of the lava flow, but it is unlikely that these produce the large plume. This is because the small jets do not provide sufficient gas volume for the larger plumes and it is difficult to develop a model in which a well-organized ~ 100 km tall plume is fed by the series of small jets spread across tens of kilometers (Milazzo et al., 2001).

Also intriguing were indications of liquid SO_2 flowing out of scarps tens of kilometers away from any active volcanism. These included small gullies seen in the highest resolution image of Io (Fig. 4) (Turtle et al., 2001) and the extensive, extremely pure SO_2 deposit seen on the floor of Baldur Patera (Smythe et al., 2000). Both Voyager and Galileo imaged scarps with morphologies highly suggestive of sapping processes (Schaber, 1982; Moore et al., 2001) (Fig. 5). However, sublimation due to solar heating is insufficient to form such scarps over geologic time (Moore et al., 1996). This leads to the conclusion that these scarps could only have formed by sapping if there were significant lateral subsurface transport of SO_2-rich fluids produced by magmatic heating. Furthermore, there is very little lag material left behind the retreating scarp (Schaber, 1982; Moore et al., 2001) (Fig. 5), suggesting that the scarp forming materials are very rich in volatiles.

2.3. Links between tectonism and volcanism

The huge tectonic massifs on Io were also a major target for Galileo observations. Perhaps some of the most useful new insights into the nature of the ionian lithosphere came from the apparent links between tectonism and volcanism.

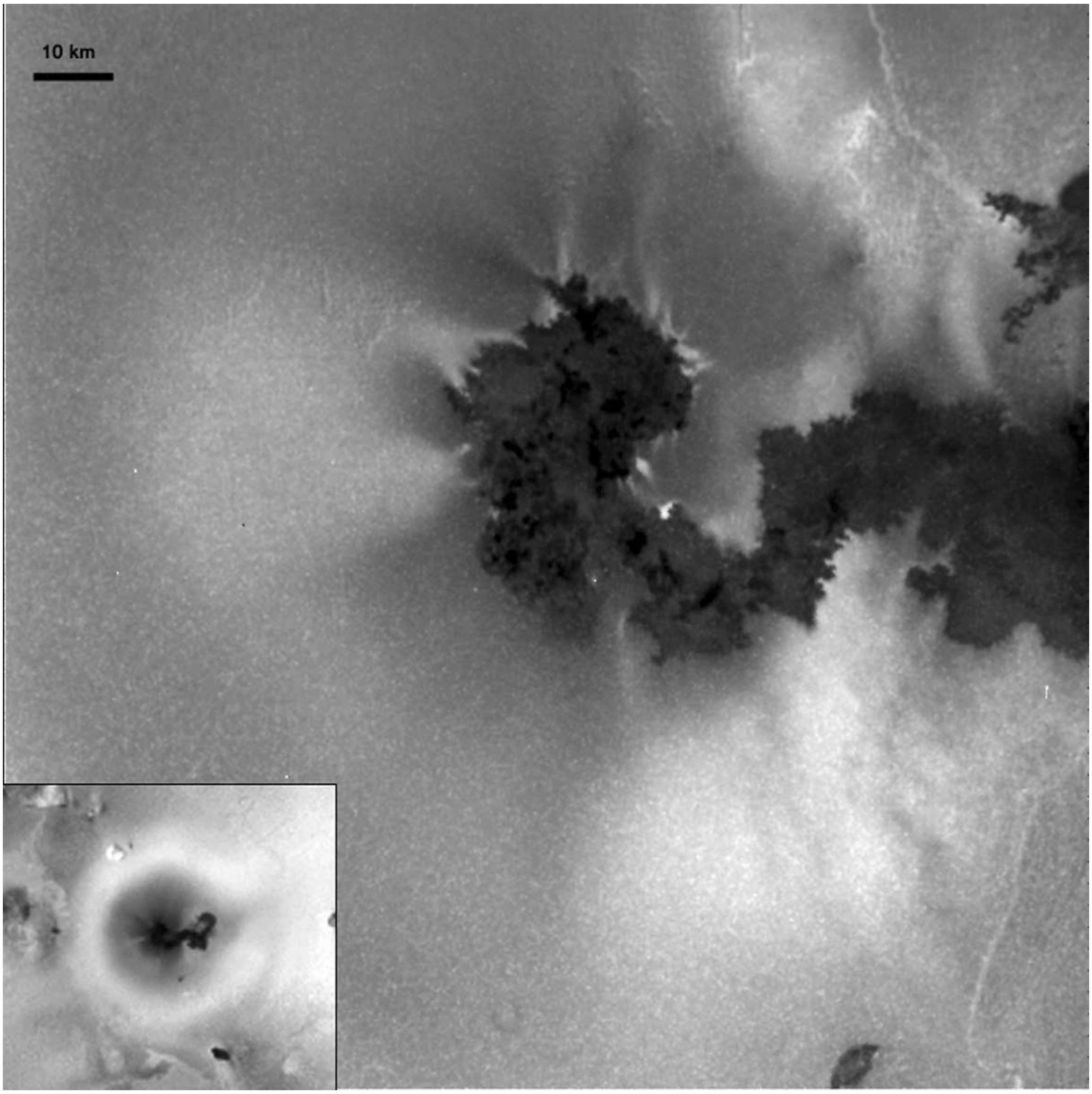

Fig. 3. SO_2-rich jets emerging from the front of active lava flows at Prometheus. Image acquired on February 22, 2000 in the green filter at 170 m/pixel with inset from an observation acquired on July 3, 1999 at 1.3 km/pixel. Color versions are available from NASAs Planetary Photojournal (http://www.photojournal.jpl.nasa.gov) as PIA-02565 and PIA-02309. The higher resolution image is of the center of the white ring of volcanic deposits seen in the inset. While it seems clear that the jets are formed by vaporization of volatiles as silicate lavas advance, the small discrete jets do not appear to create the classic umbrella-shaped plume that is centered over this same region (Milazzo et al., 2001). Instead, the jets appear to be entrained in a larger flow of gases away from the center of active flows. A larger vent, analogous to a rootless cone, may exist within the flowfield (Kieffer et al., 2000), but such a vent was not successfully imaged by Galileo. The ring deposited from the main umbrella plume extends beyond the edges of this image, as shown in the inset. North is to the top.

The Galileo data, in conjunction with the earlier Voyager images, allowed the first global inventories of mountains and volcanic centers. It was found that, on a global scale, regions with more volcanic centers tend to have somewhat fewer mountains and vice-versa (Schenk et al., 2001). However, on a local scale, the number of mountains in direct contact with volcanic depressions (paterae) is significantly higher than would be expected in a random distribution (Jaeger et al., 2003). There are also many cases where medium- to high-resolution images showed evidence that magma ascends along tectonic faults associated with mountain building (Fig. 6) (Jaeger et al., 2003).

The conclusion from these observations is that the mountains are primarily driven upward to compensate for the compression caused as materials at the surface of Io are buried, and thereby forced down to a sphere with a smaller surface area (Schenk and Bulmer, 1998; Jaeger et al., 2003). This shortening is augmented by compression due to the thermal expansion of the rocks as they approach the heated base of the lithosphere (McKinnon et al., 2001; Jaeger et al., 2003). Because the lithosphere is generally under great compression, it is not surprising that magmas preferentially rise along faults that locally relieve the compression (Jaeger et al., 2003).

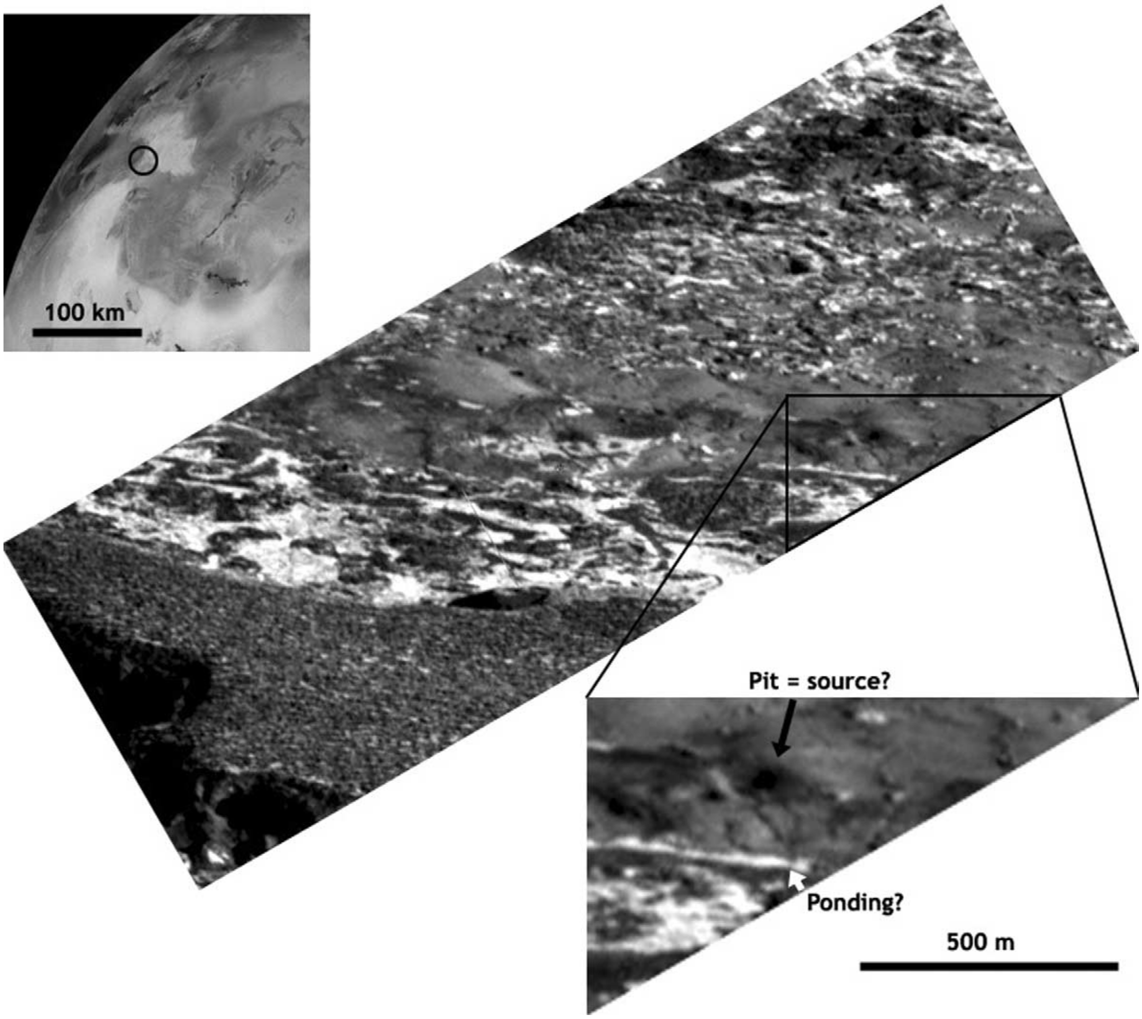

Fig. 4. Potential SO_2-carved gullies in the highest resolution image of Io. The main image is a portion of a 3-frame observation acquired on February 22, 2000 in the clear filter at 5.5 m/pixel. The view is highly oblique and the Sun is to the right (East). The entire observation, reprojected to an overhead view is available at the NASA Photojournal as PIA-02562. The inset is the best available context at 1.3 km/pixel with the general region of the high-resolution observation circled. North is approximately to the top of the figure. The gullies appear to connect a pit surrounded with darker diffuse deposits to a layer of bright material hugging the base of a scarp. One possible interpretation is that the pits are formed by SO_2 liquid geysering to the surface as proposed by McCauley et al. (1979), scattering darker non-volatile material around the vent. In this scenario, liquid SO_2 has cut into an unconsolidated surface and has ponded on flat topography. If the frozen SO_2 is meters thick, it could take hundreds of years to sublimate (Moore et al., 1996). The partially imaged mesa in the lower left of the Galileo image is ~ 400 m tall. If the surrounding variegated and hummocky terrain is the lag from scarp retreat, it suggests that the mesa material is dominantly volatiles.

2.4. Geophysics

The geophysical measurements from Galileo provided some of the most illuminating data on the interior of Io. The gravity field of Io has been modeled using both a pure iron and an Fe–FeS eutectic composition (Anderson et al., 2001). However, based on the fact that Io is seen to have significant sulfur, the Fe–FeS composition was strongly favored. The radius of the core should be slightly less than half the radius of Io. The gravity data led to the conclusion that the L and LL classes of meteorites are the only reasonable candidates for the bulk composition of Io (Anderson et al., 2001). However, it is important to note that the L and LL chondrite compositions would not provide enough sulfur for the core to be at the Fe–FeS eutectic. More recently, alternative models with elemental ratios closer to solar abundances and a sulfur and oxygen-rich core have been used to explain the bulk density of Io (McKinnon and Desai, 2003). The density structure of the mantle and crust is less well constrained, but relatively low densities for the crust and upper mantle are favored.

The magnetometer experiment on Galileo determined that Io has no significant intrinsic magnetic field (Kivelson et al., 2002). This observation suggests (but does not prove) that Io's core is not convecting, since motions within the conductive core are thought to be the key to geodynamos (e.g., Nimmo, 2002). The lack of convection is consistent with all proposed models for tidal dissipation within Io because in all of these models the heating of the mantle prevents the core from cooling.

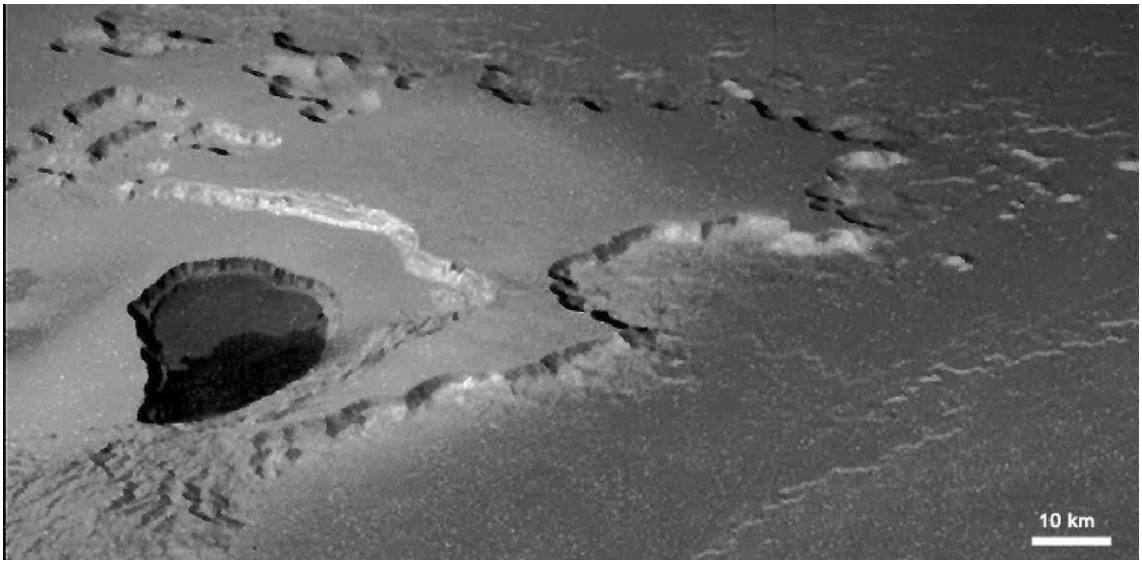

Fig. 5. Apparent sapping scarps at Tvashtar Catena. Moderately oblique image acquired on November, 25, 1999 through the clear filter at 180 m/pixel. North is approximately to the top of the image. The flat-topped mesa is about 1 km tall and exhibits classic sapping morphologies (Moore et al., 2001). The thin lag deposits suggest that the bulk of the mesa is volatile material. However, the more rugged mountain to the southwest appears to be less volatile rich. Heat for mobilizing the presumably SO_2-rich volatiles would come from the active volcanism within Tvashtar Catena, not from solar illumination. The dark-floor of the inner depression contains recent silicate lavas and an active curtain of lava was erupting just to the west of this image. Colorized version of the entire observation is available as NASA Planetary Photojournal PIA-02545.

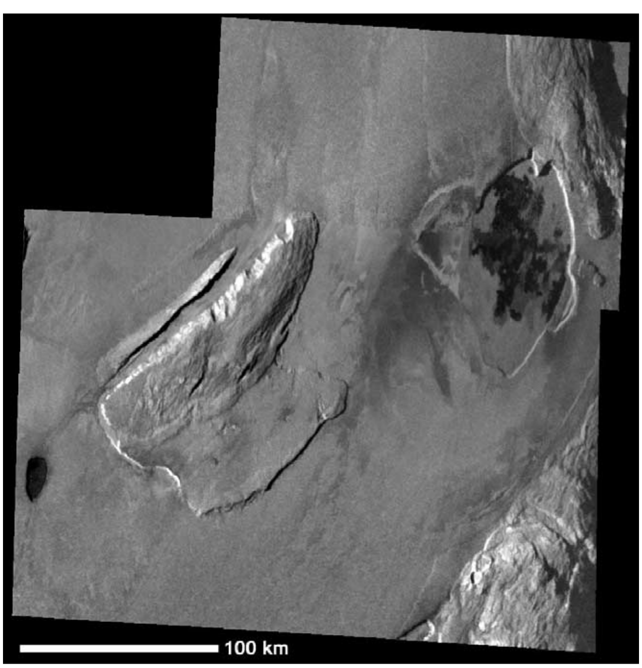

Fig. 6. Possible links between tectonism and volcanism. This Galileo observation of Shamshu Mons and Patera was taken on February 22, 2000 through the clear filter at 345 m/pixel resolution. North is toward the top of the mosaic. The smaller dark-floored patera on the west end of the mosaic lies on the fracture system that breaks the front of the mountain in the center of the mosaic. The larger patera to the northeast is in direct contact with a mountain, in a region where the mountain uplift could have locally relieved compressional stresses (see discussion in Section 2.3). These types of observations suggest that magma preferentially rises through the lithosphere along major fault systems (Jaeger et al., 2003).

3. A post-Galileo model for the interior of Io

3.1. Core

It may seem peculiar, but Io's core is no worse understood than the rest of Io's interior. The composition is most likely to be a mix of Fe and S. If the core composition were near that of the Fe–FeS eutectic, then it would have a density of $\sim 5150\ \mathrm{kg/m^3}$ and a melting temperature ~ 1200–1300 K (Brett, 1973; Anderson et al., 2001). We expect the base of the mantle to be significantly hotter than this. In fact, in the following section we suggest that the base of the mantle is likely to be near the melting temperature of pure Fe (~ 1900 K at these pressures ~ 3 GPa). Therefore, it is reasonable to expect Io's core to be completely molten, consistent with the magnetometer data, even if only a small amount of sulfur is in the core. The radius of the core should be between 550 and 900 km, with the larger core favored by lower-density mantle compositions (Anderson et al., 2001). If the core contains significant oxygen, it could be as large as 1100 km in radius (McKinnon and Desai, 2003).

3.2. Mantle

If we accept the conclusion that Io has an L or LL ordinary chondrite bulk composition, and an Fe-rich core, we can estimate the bulk composition of the remaining silicate mantle. Table 1 shows our best estimate for the major element composition of the silicate portion of Io. The relatively high MgO and SiO_2 contents that we estimate require that enstatite be the most common mineral in the mantle.

Table 1
Estimated Bulk Composition of Io's silicate portion

Wt%	L-chondrite	LL-chondrite	Earth
SiO_2	54.0	53.0	48.0
TiO_2	0.16	0.17	0.27
Al_2O_3	3.1	2.9	5.2
FeO*	5.0	7.8	7.9
MgO	34.0	33.0	34.0
CaO	2.5	2.5	4.2
Na_2O	1.3	1.2	0.33

The estimated bulk composition of Earth's silicate portion (Anderson, 1989) is included for comparison. Io's silicate composition is estimated by removing 10 wt% FeO (representing the core) from the bulk Io compositions from Kuskov and Kronrod (2001).

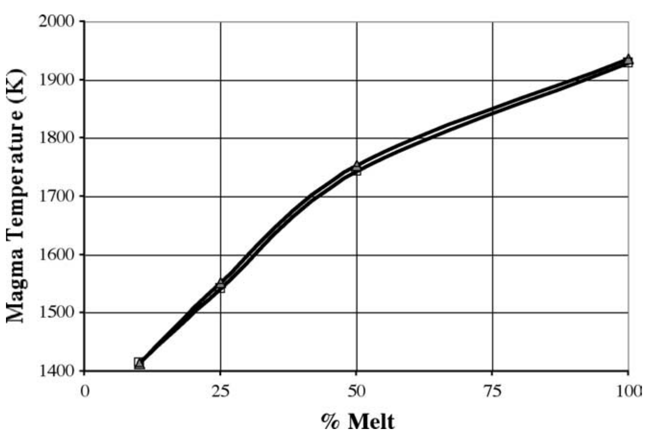

Fig. 7. Estimated melt fraction in Io's mantle as a function of lava temperature. Temperatures were calculated using the MELTS program (Ghiorso and Sack, 1995) and the bulk compositions listed in Table 1 at 100 MPa and an oxygen fugacity at the iron–iron wustite buffer. This pressure corresponds to about 20–30 km depth, which is our best estimate of the lithospheric thickness (see discussion in Section 3.3). The results are not sensitive to the choice of oxygen fugacity or the differences in the compositions from the L and LL chondrite models. Results using the L chondrite composition are plotted with shaded triangles and the LL composition with open squares. The highest temperatures seen on Io correspond to about 80% melting of the mantle but temperatures corresponding to 30–60% melting are more commonly seen. This suggests that the top of Io's mantle is roughly half solid and half liquid.

While it is relatively straightforward to derive a rough estimate of the bulk composition of the mantle, its state (molten, mushy, or solid) cannot be directly determined from the available observations. However, if we take the compositions and temperatures of the erupted lavas derived from the Galileo data at face value, we do have hard constraints on the magma source region (i.e., the mantle). We used the MELTS program (Ghiorso and Sack, 1995) to calculate the degree of melting at different temperatures (Fig. 7). MELTS is a numerical thermodynamic model that combines published data from > 2500 laboratory experiments on a variety of silicate compositions at many pressures and oxygen fugacities. We concentrate on a pressure of 100 MPa, since this corresponds to a depth of 20–30 km in Io, which is our best estimate for the thickness of the lithosphere (e.g., Section 3.3; Carr et al., 1998; Jaeger et al., 2003). We find that, in order to produce the hottest lavas seen on Io ($\geqslant 1870$ K), the ionian mantle must be locally $\geqslant 80\%$ molten. Color temperatures of 1500–1600 K were observed frequently (McEwen et al., 1998b), which implies a magma temperature of ~ 1800 K (Keszthelyi and McEwen, 1997). This suggests that $\sim 50\%$ partial melting of the ionian mantle may be quite common. These lavas would be $\sim 50\%$ magnesium-rich orthopyroxene, $\sim 30\%$ plagioclase, $\sim 15\%$ clinopyroxene, and 5% other minerals. Such a mix of minerals is an excellent match to the SSI color data. Moreover, $\sim 50\%$ partial melting may also be sufficient to disaggregate the crust at its base allowing it to be efficiently mixed back into the mantle, as required to avoid differentiating Io.

If no chill crust formed on the ionian lavas and the color temperatures of ~ 1500 K corresponded to the actual magma temperatures, then only $\sim 25\%$ melting would be required. However, these lower temperature lavas would be dominated by plagioclase and clinopyroxene, rather than the orthopyroxene indicated by the SSI color data. We also find 100% melting to be unlikely because the lavas would be expected to contain substantial olivine, which is not seen in the SSI color data. Therefore, in our model we suggest that the top of Io's mantle is about half solid and half liquid. Such a layer would constitute a "global mushy magma ocean" as described in Keszthelyi et al. (1999). This magma ocean seems to be required by the Galileo data unless

(1) there is an exotic mechanism to superheat the lava,
(2) the SSI color data have been misinterpreted, or
(3) Io has a bulk composition significantly different from any chondritic meteorite.

We use our model's global layer of $\sim 50\%$ molten rock to extrapolate to the deeper mantle. If we assume that Io's mantle is convecting, the temperature profile should be roughly adiabatic. It is important to note that numerical modeling of convection within Io by Tackley et al. (2001) suggests that Iotherms could be somewhat sub-adiabatic. In these models, the temperature at the core–mantle-boundary could be as much as 200 K lower than would be predicted by an adiabatic temperature gradient. Using MELTS, we can follow the phase changes that occur along the adiabat (i.e., as pressure increases and entropy is conserved) and along an isothermal temperature gradient that approximates the Tackley et al. (2001) model results. Figure 8 shows the changes in crystallinity and temperature with depth within Io's mantle predicted by the MELTS model. Both adiabatic and isothermal temperature profiles predict that there will be substantial melting throughout the mantle, with 10–20% melt at the core–mantle-boundary. Thus our new model is broadly similar to the "thick asthenosphere" model preferred in the pre-Galileo models (e.g., Spencer and Schneider, 1996), but involves substantially higher degrees of partial melting than had been previously envisioned. At the same time, the crystallinity is much higher than the Voyager-era magma ocean

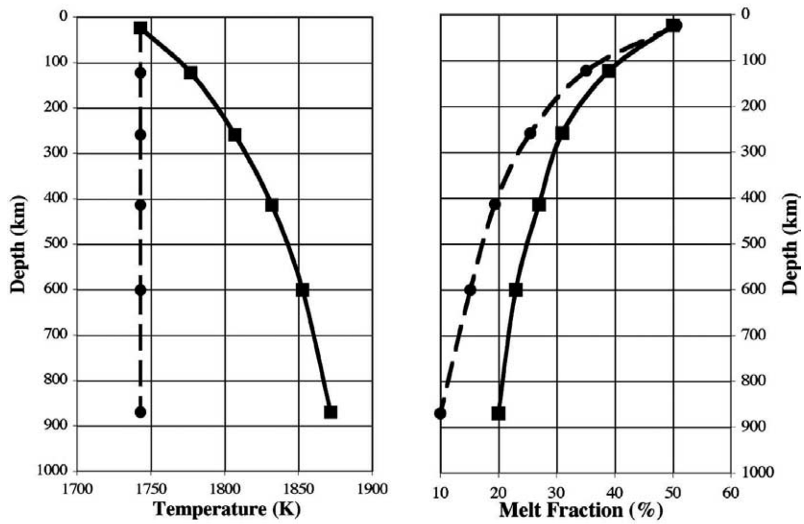

Fig. 8. Plausible temperature and melt profiles through the ionian mantle. Modeled using the MELTS program (Ghiorso and Sack, 1995), assuming the LL chondrite composition from Table 1, starting with 50% melting at the top of the mantle, and descending along an adiabat (solid line with squares) or isotherm (dashed line with circles). The melt fraction is predicted to monotonically decrease with depth, reaching 10–20% at the core–mantle boundary. In the adiabatic case the mantle temperature gradually increases from near 1700 K at the top to near 1900 K at the base of the mantle. In the isothermal case, it is assumed that convection is so efficient that there are no significant temperature variations within the mantle as suggested by some model runs in Tackley et al. (2001). In either case, density is predicted to smoothly increase from slightly less than 2900 kg/m^3 to just above 3100 kg/m^3 from the top to the bottom of the mantle.

models. We will discuss the physical plausibility of this model in Section 4.1.

3.3. *Lithosphere/crust*

The Galileo observations of the links between volcanism and tectonism have supported the lithospheric model of O'Reilly and Davies (1981). The rapid volcanic resurfacing pushes cold crust downward and mountains are forced up to compensate (Schenk and Bulmer, 1998). In order to explain the observed mountains, the lithosphere must be at least 13 km thick (Jaeger et al., 2003). It is also geometrically impossible to produce the observed mountain dimensions if the lithosphere is > 80 km thick (Jaeger et al., 2002). Galileo-era researchers (Carr et al., 1998; Jaeger et al., 2003) prefer a lithospheric thickness of 20–30 km.

Figure 9 shows the predicted temperature within the lithosphere away from active volcanic centers and tectonic faults. Intrusions of hot magma will perturb the average temperature profile, but we expect such perturbations to be highly localized and be volumetrically minor (Jaeger et al., 2003). This is because

(a) volcanic vents make up a very small fraction of Io's surface, and
(b) thermal conduction is slow compared to the estimated rate of crustal recycling.

While warm lavas with albedos of 0.2–0.3 make up 1–2% of Io's surface, the darkest surfaces that correspond to the most recently active silicate lavas (as opposed to stagnant cooling flows) make up only on the order of 0.1% of the surface (McEwen et al., 1985). The area of the actual vents for these lavas is probably even smaller than that of the darkest lavas, so the plumbing system for these lavas should make up $\ll 1\%$ of Io's crust. Even if the plumbing system were active for 10^6 years, the Iotherm would be raised by > 100 K only within 10 km of the active magmatic system (assuming a thermal diffusivity $\sim 10^{-6}$ m^2/s). This is likely to be the maximum length of time a magmatic plumbing system can survive because the entire lithosphere is expected to be recycled on this time scale (e.g., Johnson et al., 1979). Thus, thermal perturbations should be localized to within a few kilometers of the actual magma conduits. However, it is important to consider that fact that there may be more intrusions in older parts of the crust. This means that while heat from magmatic activity will be inconsequential for $\gg 90\%$ of the ionian upper crust, it may be non-trivial in the deeper, older, portions of the crust.

An active "hydrothermal" system involving liquid SO_2 could elevate Iotherms more regionally. The presence of mobile liquids in the shallow subsurface is indicated by the observations of SO_2-rich flows (e.g., Smythe et al., 2000; Williams et al., 2004; Turtle et al., 2004) and sapping scarps (e.g., Moore et al., 2001). These fluids should fill any empty pore spaces below the local "water-table." Therefore, we expect a sharp jump in the density of the crust at this depth, as proposed by Leone and Wilson (2001). An Iotherm consistent with active circulation of liquid SO_2 is plotted in Fig. 9, however, it is likely that a liquid SO_2 "aquifer" is not global in extent and is instead limited to specific regions of Io (Moore et al., 2001).

With either Iotherm, the bulk of Io's lithosphere is extremely cold and the silicate portion will behave very rigidly. Also, if the crust is defined on the basis of the difference in composition between the lavas and the mantle of Io, then

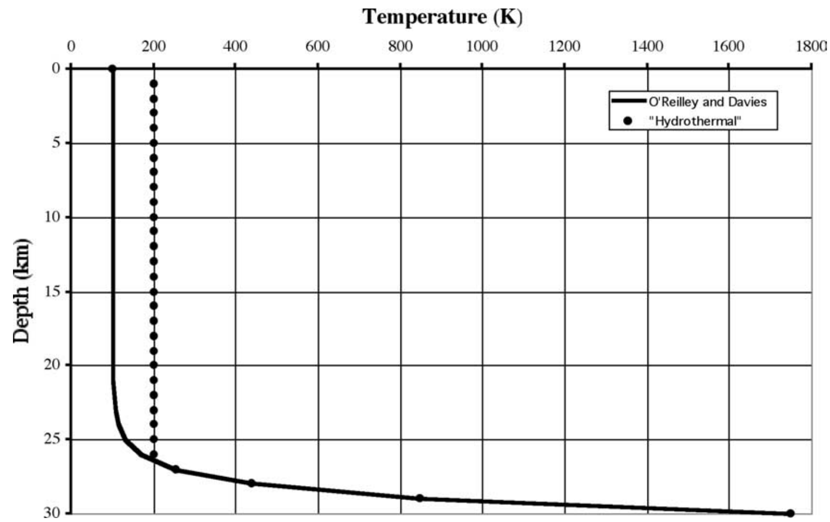

Fig. 9. Plausible temperature profiles through the ionian crust, away from magmatic activity. The solid line is the temperature profile from O'Reilly and Davies (1981) assuming a resurfacing rate of 1 cm/yr and a lithospheric thickness of 30 km. The dotted line is representative of a temperature profile buffered by the freezing and melting of SO_2. However, the phase diagram for SO_2 is very poorly known due to a lack of experiments at higher pressures, so it is plausible that this line should become somewhat hotter with depth. In either case, sulfur will melt only in the bottom few kilometers of the crust, and the bulk of the silicates in the crust will be extremely cold and rigid. These temperature profiles are likely to apply to most of the young upper crust. However, more of the older lower crust is likely to have been perturbed by magmatic activity.

there is a very close correspondence between the crust and the mechanical lithosphere. Only in its lowermost few kilometers is the crust warm enough to be mechanically weak over geologically interesting time scales.

4. Discussion

While the above model for the interior of Io remains unproven, it is consistent with the available observations. At this point, it seems important to discuss two key questions:

(1) is this model physically viable, and
(2) what are the implications if this model is accurate?

4.1. Can an ionian magma ocean exist?

The most controversial aspect of this model is the mushy magma ocean. A purely liquid magma ocean has been ruled out because it would make it very difficult to maintain the tidal heating and topography seen on Io (e.g., Ojakangas and Stevenson, 1986; Webb and Stevenson, 1987). A mushy magma ocean has been considered "implausible" because it has been assumed that the melt would escape from the mantle before such high degrees of partial melting could be achieved. In fact, recent modeling of the thermal and dynamic processes within Io suggests that $< 20\%$ partial melting is needed to achieve an energy balance between tidal heating and volcanism (Moore, 2001). Io is expected to have a negative feed-back loop, where increased melting leads to even faster heat loss and thus the mantle is predicted to be cold and largely solid (Ojakangas and Stevenson, 1986; Fischer and Spohn, 1990; Moore, 2001; Monnereau and Dubuffet, 2002).

The recent thermo-physical models indicating a largely solid Io (Moore, 2001; Monnereau and Dubuffet, 2002) stand in stark contrast to the geochemical modeling (Keszthelyi and McEwen, 1997) that indicates that 10–25% partial melting would result in ionian lavas very different from what we observe. It is important to examine whether there is any reason to question some aspects of the recent thermo-physical models. In particular, does the statement that 20% partial melting is sufficient to explain Io's heat flow (Moore, 2001) really preclude higher degrees of partial melting? Since Carr et al. (1998) calculate that 20% partial melting is actually the minimum needed to explain Io's volcanism, this question deserves some discussion.

If, despite a high degree of partial melting, the magma generated within Io is only able to escape slowly (i.e., at the rate that corresponds to $\sim 20\%$ partial melting in the models), then there is no conflict between the recent thermo-physical and geochemical models. Thus, the plausibility of the mushy magma ocean rests on the plausibility of mechanisms to retard the extraction of magma from within Io. There are several reasons to suggest that magma extraction may not be as efficient as in the models.

Neither the work of Monnereau and Dubuffet (2002) nor most of the earlier studies directly examine melt extraction. However, Moore (2001) provides a detailed physical model for melt migration based on porous (Darcy) flow of a buoyant fluid through a solid matrix. The relative velocity between the liquid and the solid (v) is given by

$$v = \left(b^2\phi^n\Delta\rho g\right)/(c\phi\eta),$$

where b is the typical grain size, ϕ is the melt fraction, $\Delta\rho$ is the density contrast between the melt and the solid, g is the gravitational acceleration, η is the melt viscosity, and n and c are constants determined by the geometry of the melt pathway (Moore, 2001). Despite many simplifying assumptions, these equations have been used extensively to describe melt extraction from the terrestrial mantle (e.g., McKenzie, 1985; Turcotte and Schubert, 1982; Stevenson and Scott, 1991). The most significant limitation of this model is that it is not valid once the solid matrix begins to disaggregate. This process is often assumed to happen at $\sim 40\%$ partial melting, but there are situations where a solid network can form even at $\sim 80\%$ partial melt (Philpotts et al., 1998). Thus, this equation is plausibly (but not necessarily) valid even for a mushy magma ocean.

More challenging is the selection of values for the different parameters in the equation. For the values favored by Moore (2001) ($b = 1$ cm, $\Delta\rho = 500$ kg/m^3, $g = 1.8$ m/s^2, $\eta = 1$ Pa s, $n = 3$, and $c = 200$), v is 1.8×10^{-5} m/s at 20% partial melting. If one assumes, as did Carr et al. (1998), that the typical crystal size is ~ 1 mm, then v is 1.1×10^{-6} m/s even at 50% partial melting. Thus, if Io's mantle has reasonably small crystals, melt extraction rates could be slower than suggested by Moore (2001) even with 50% partial melting. There are similarly significant uncertainties in the viscosity of the melt and the density contrast between the melt and the solid. Our MELTS modeling suggests that the density contrast varies between 520 and 690 kg/m^3 and melt viscosity varies between 1 and 400 Pa s as partial melting varies between 50 and 20% (more melting leads to lower density contrasts and lower viscosity liquids). Thus, it is plausible (although by no means certain) that the observed volcanic heat flux is consistent with 50% partial melting.

There are additional reasons why the efficiency of magma extraction may have been overestimated in earlier studies. It is generally assumed that the lithosphere does not provide a significant barrier to magma ascent (Ojakangas and Stevenson, 1986; Moore, 2001; Monnereau and Dubuffet, 2002). However, the tectonic features on Io indicate that the lithosphere is under extreme compression (Schenk and Bulmer, 1998; Jaeger et al., 2003). This situation appears to restrict magma ascent through the lithosphere to locations where faulting has locally relieved the compression (Jaeger et al., 2003). Thus the rate-limiting process for the eruption of lavas may be the tectonics in the lithosphere, not magma migration in the mantle.

There are also questions about how much heat could be generated in a mantle that is weakened by large degrees of partial melting. At low degrees of partial melting, the two-phase mantle can be approximated as a viscoelastic solid. In such a material, tidal energy dissipation peaks at a few percent partial melting and rapidly drops with increasing melt fraction (e.g., Segatz et al., 1988; Fischer and Spohn, 1990). However, at large melt fractions viscous dissipation within the liquid can be significant (Ross and Schubert, 1985). This process can dissipate energy very efficiently if the tides push the liquid back and forth through a solid matrix (Stevenson, 2002). However, the complex behavior of this type of two-phase material will require extensive new modeling and experimental efforts (Stevenson, 2002). Such future work may show that the mushy magma ocean model will need to be further refined, or even rejected.

There are additional unresolved questions about the validity of the mushy magma ocean model that await future research. For example, it is unclear how mix of crystals and melt would avoid physical separation into a purely solid interior and a purely liquid exterior. Stevenson (2002) predicts that a mush zone > 20 km deep would be unstable over geologic timescales. Another issue is that, if the temperature of the mantle were to change significantly on a time scale of less than 10^6 years, then our model for stresses in the lithosphere would be inaccurate (McKinnon et al., 2001). It is even possible to construct a model, with heating from the top and realistic temperature dependent physical properties, in which Io's mantle is not convecting and has a cold, completely solid, interior. These issues show that interior of Io will require continued investigation before any definitive conclusions can be drawn. However, the model we have presented is consistent with the available observations and has not (yet) been demonstrated to be physically implausible. As such, it can serve as a useful starting point for future discussions.

4.2. Consequences for surficial processes

Irrespective of our model for Io's mantle, our crustal model has some interesting implications for the formation of one of the geomorphologic features unique to Io: paterae. Paterae on Io are irregular depressions typically filled with lava. While superficially similar to terrestrial calderas, their formation has been difficult to reconcile with caldera formation mechanisms (Radebaugh et al., 2001). In particular, both silicic and basaltic calderas are formed by syn- or immediately post-eruption collapse of the roof of a shallow magma chamber. In the case of paterae, there is no evidence for sufficiently voluminous eruptions outside the paterae to account for the volume of magma required to produce the observed depression. One possibility is that paterae are largely tectonic depressions that have been modified by volcanic processes taking advantage of local extension (Radebaugh et al., 2001). While tectonic processes must play an important role in the formation of those paterae that are associated with the fault systems surrounding mountains, there are many paterae with no obvious nearby tectonic activity. Silicate magmas ascending through our hypothesized crust of Io lead to another formation model for paterae (Fig. 10).

Regions with higher volcanic activity are likely to have enhanced mobility of SO_2 within the subsurface, leading to an enrichment of volatiles near the top of the crust. Ascending silicate magmas would not be able to buoyantly rise through this low-density zone, and would have the tendency to form sills at the base of the zone. The heat from this sill

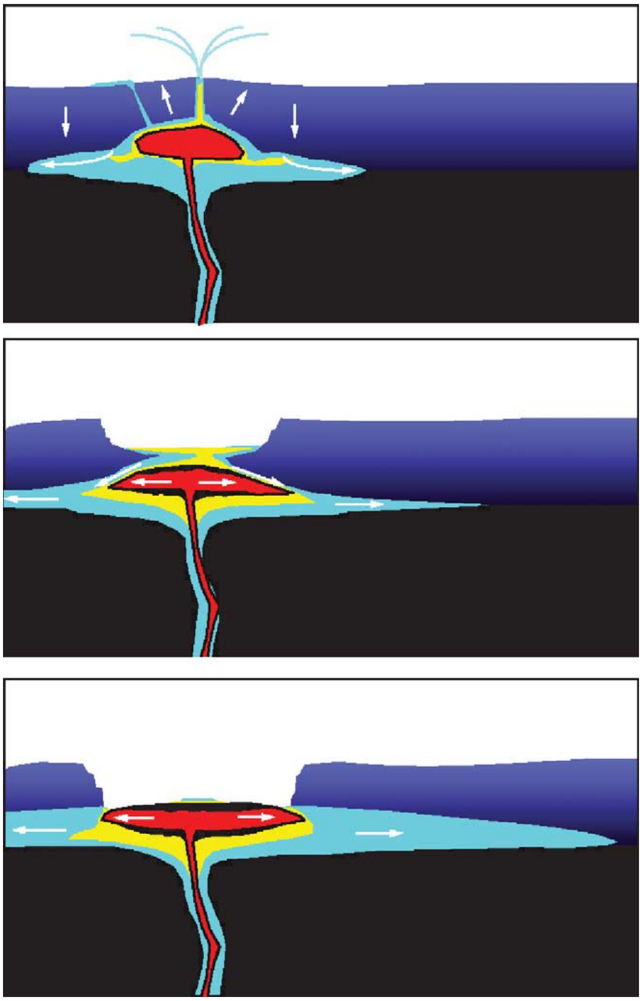

Fig. 10. Cartoon of patera formation model. Silicate magma (red) is able to ascend through solid rock (black) but lacks the buoyancy to rise through a volatile rich zone (dark blue) near the surface, especially if this zone is porous. The heat from the intrusion melts sulfur (yellow) and SO_2 (light blue) which migrate into the permeable surroundings. Where SO_2 is vaporized, it will erupt to the surface, sometimes entraining sulfur with it. As this process continues, a substantial depression can form over the intrusion. If the flux of silicate magma is sustained, it may become locally, or even completely, unroofed.

would initially melt the SO_2-rich volatiles, allowing them to flow away in the subsurface. This flow could remove a substantial volume of the overlying volatiles and might only be seen at the surface as sapping scarps. As the materials surrounding the sill are heated further, SO_2 will vaporize and then sulfur will melt. At temperatures approaching those of the silicate magmas, sulfur will also vaporize. Therefore this model predicts the eruption of light colored plumes rich in SO_2, yellowish sulfur flows, and red S-rich plume deposits around the growing patera. Such deposits are indeed seen around many paterae (Williams et al., 2002, 2004) (Fig. 11).

If this activity persists long enough, the sill will be unroofed, exposing the silicate lava. The sill will have been expanding laterally and a cooled crust should have grown over the oldest (i.e., central) part of the sill. This may help explain the islands of volatile-rich materials seen in the centers of some paterae (e.g., Fig. 12). As the silicate lavas continue to volatilize the edges of the patera, some sulfur-rich liquids are expected to flow over the solidified silicate lavas. Such intermingling of silicate and sulfur-rich lavas is seen in many paterae. The continued lateral expansion of the lava would also constantly undermine the walls of the patera. Such an active process would help explain the steep walls observed to surround many paterae (e.g., Figs. 5, 6, 11, and 12). However, it is important to note that the conclusion of Clow and Carr (1980) that the steep walls surrounding paterae could not be sulfur-rich is dependent on the assumption that high conductive heat flow through the crust would melt sulfur just a few hundred meters under the surface. Given the cold temperatures we now expect within the crust, the same analysis by Clow and Carr (1980) predicts that sulfur-rich material could support vertical scarps > 10 km tall.

We can use the predicted timescale over which a sill can exhume itself to provide a test of the plausibility of this patera formation model. We do this via a simple heat conduction model. This model ignores phase changes and advection of the volatiles, but is sufficient for an initial, crude plausibility test. If we assume that the sill is constantly recharged and consider only 1 dimension, we can use the solution for heat conduction in an infinite half-space with one boundary held at a fixed temperature:

$$T = T_o + \left((T_m - T_o) * \operatorname{erfc}\left(z/2\left((Kt)^{1/2}\right)\right)\right),$$

where T_o is the initial temperature, T_m is the magma temperature, z is the distance from the sill, K is the thermal diffusivity, and t is time. Reasonable inputs are $T_o = 100$ K, $T_m = 1600$ K, and $K = 3 \times 10^{-7}$ m^2/s for the presumably somewhat porous, volatile-rich layer. SO_2 will melt if the temperature rises by ~ 100 K, and sulfur will melt if the temperature rises by ~ 300 K. A kilometer from the sill, it will take $\sim 16{,}000$ years for SO_2 to begin to melt and $\sim 33{,}000$ years for sulfur to begin to melt. One hundred meters from the sill, the melting begins in 160 and 330 years for SO_2 and sulfur, respectively. Since most paterae are hundreds of meters to a few kilometers deep, this suggests that patera formation typically takes on the order of 10^2–10^5 years. This is consistent with both the estimate that Io's surface is younger than 1 Ma (e.g., Johnson et al., 1979; Phillips, 2000) and the fact that no new paterae have formed over the 20 years spanned by Voyager and Galileo (e.g., McEwen et al., 1997). This simple calculation justifies the more complete analysis of the heat and mass transport around a sill intruded into volatiles that we are initiating.

5. Conclusions

Our immediately post-Galileo model of Io is, in some ways, more like the immediately post-Voyager view than the immediately pre-Galileo view. In particular, we suggest that the interior of Io includes a large volume of liquid. The

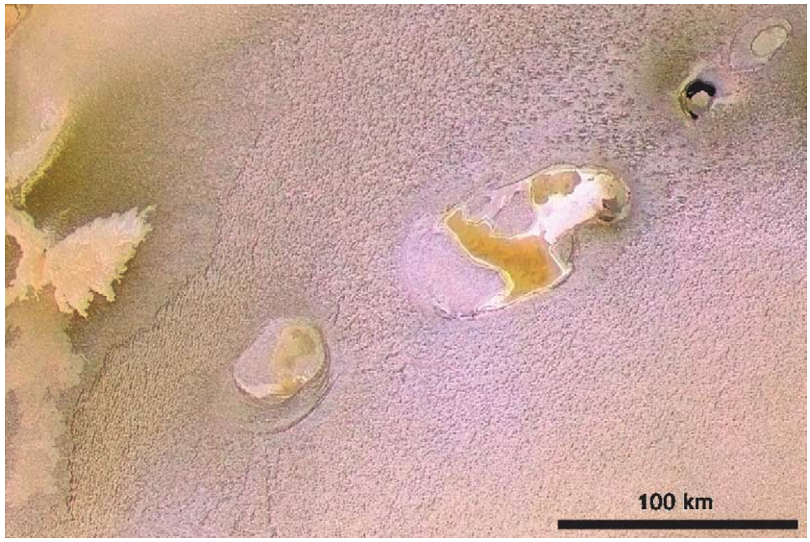

Fig. 11. Portion of false color mosaic over the Chaac–Camaxtli region of Io. This observation was acquired on February 22, 2000 through the clear filter at a resolution of 185 m/pixel and combined with 1.3 km/pixel color observation collected on July 3, 1999. The image uses the violet, green, and 756 nm filters, producing colors slightly redder than would be apparent to the human eye. North is to the top. The full observation is available from NASA Photojournal as PIA-02566. Note the eruption of light-colored lavas near the west edge of the figure. These lavas are interpreted to be SO_2-rich fluids mobilized by shallow silicate intrusions. The small dark spot and bright red diffuse deposit at the source of the bright lava flows indicate that a very small amount of silicate and/or sulfur lava has also erupted. This would represent the first stage in our new model for patera formation. The margins of the southernmost and easternmost paterae appear to have relatively shallow slopes, as could form over a region where volatiles (again mobilized by shallow silicate intrusions) were able to migrate laterally through a porous upper crust. These morphologies would form in the second stage of the patera formation model. The poorly resolved ridges on the plains could be formed by deformation of the surface over areas with high subsurface flow. The largest patera shown has bright (white) and yellow lavas on its floor. This coloration suggests that substantial volumes of molten sulfur, as well liquid SO_2 are available, possibly because this patera has undergone more sustained heating from a silicate intrusion. The small, dark-floored patera to the northeast may represent the final stage of patera formation where the silicates have been unroofed by the removal of the overlying volatile rich layer.

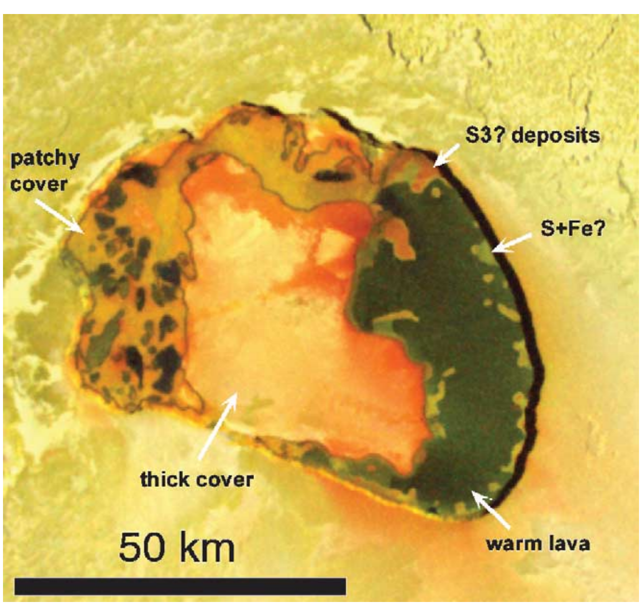

Fig. 12. Medium-resolution, visible observation of Tupan Patera. SSI observation acquired on October 16, 2001 at a resolution of 135 m/pixel with the data from the violet, green, and 756 nm filters displayed. North is to the top. NASA Planetary Photojournal PIA-02599. Simultaneous NIMS data show that the dark, eastern region within the patera is the warmest, but the western and northern mottled floors are also warm. The light colored central "island" is cold (Lopes et al., 2004). Yellowish and whitish surfaces are interpreted to consist of SO_2 tainted with various amounts of elemental sulfur. Orange surfaces may be relatively pure sulfur. Red diffuse deposits are thought to be colored by metastable, short-chained sulfur produced by condensation of hot sulfur vapors. Green surfaces appear to form by reaction between warm, dark silicate lavas and the red diffuse deposits (McEwen et al., 2003; Williams et al., 2004), possibly involving the creation of iron sulfur compounds. This distribution of silicate and sulfurous materials matches the last stage of our patera formation model. Tupan may be an example of a silicate sill that has largely exhumed itself and is actively melting and vaporizing the patera walls. The mottled appearance of part of the floor appears to be due to liquid sulfur flowing from the walls of the patera onto a slightly irregular, warm silicate surface. This active erosion of the patera walls may be an essential ingredient in maintaining the near vertical, ~ 900 m tall cliffs. The central "island" may be an area where the removal of volatiles is slowed by the presence of a thick, stable, insulating crust on the top of the older part of the sill.

iron- and sulfur-rich core is expected to be about half the radius of Io and molten. The mantle is expected to range from $\sim 10\%$ partial melt at its base to as much as $\sim 50\%$ at its top. The ascent and eruption of the mantle melts may be retarded by the strength of the solid portion of the mantle, the compressive stresses in most of the lithosphere, and a volatile-rich low-density layer in the uppermost crust. Heat from magmatic intrusions may provide the energy to mobilize volatiles in Io's subsurface, helping to produce features like sapping scarps and paterae. Io remains an extremely puzzling and dynamic world, but the model we present is consistent with the available observations and does not (as yet) appear to violate known physical and thermodynamic constraints. We invite vigorous testing of our model.

Acknowledgments

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