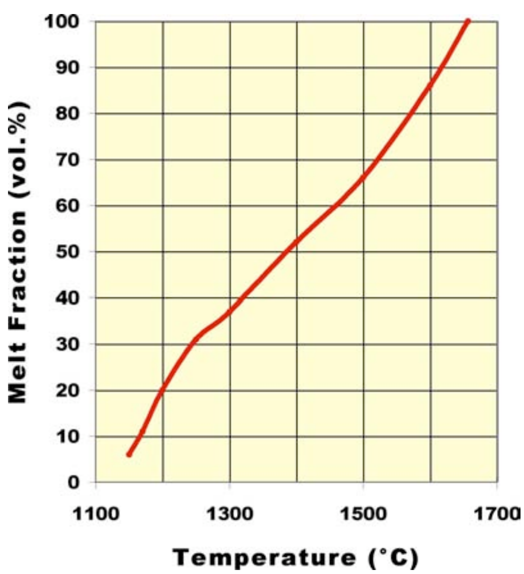
RECONCILING LAVA TEMPERATURES AND INTERIOR MODELS FOR IO. L. Keszthelyi[1], M. Milazzo[2], W. L. Jaeger[2], L. Wilson[3], and K. L. Mitchell[3], [1]Astrogeology Team, U.S. Geological Survey, Flagstaff, AZ 86001, USA, [2]Lunar and Planetary Laboratory, University of Arizona, Tucson, AZ 85721, USA, [3]Environmental Science Department, Lancaster University, Lancaster, LA1 4YQ United Kingdom.

Introduction: The very high temperatures that have been reported for lavas on Io [1-3] have proven to be difficult to explain [4]. In particular, the estimates of lava temperatures of ~1600°C require that the interior of Io be (a) largely undifferentiated and (b) almost completely molten (Fig. 1). However, models of tidal dissipation within Io show that a largely molten interior would not produce sufficient heat to maintain the observed level of volcanism [5-8]. The latest modeling suggests that Io's mantle should not contain more than 20% melt [7]. If Io's bulk composition is similar to a CM chondrite [9-11], then 20% melt implies a lava temperature of 1200 °C [11] (Fig. 1). This 400°C discrepancy is the focus of this study.

Figure 1. Melting curve for Io. Composition from [11]. Computed using the pMELTS program [12] at 0.2 GPa and fO_2 set at QFM [13].



Revised Temperature Estimates: One avenue of investigation has been to improve our error analysis of the temperature estimates. We look at two locations (Pele and Tvashtar) where Galileo SSI observations resolved the hot spots and the data suggested ultramafic temperature (>1300°C).

The improved error analysis hinges on the recognition that the noise distributions in SSI images of Io are not "normal" or Gaussian. Instead, the noise distribution is jagged with a long tail to high DN values (Fig. 2).

Figure 2. Histogram of DN values from dark area of high resolution I32 observation of Pele. Blue and purple curves are from a clear and 928 nm filter image, respectively.

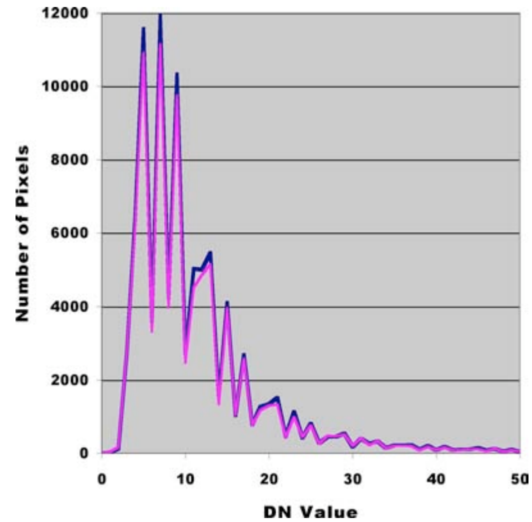
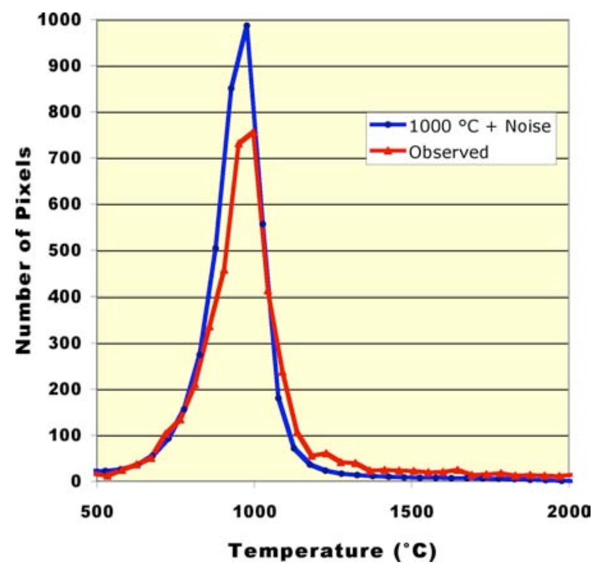


Figure 3. Modeled and observed temperature distributions at Pele. Model uses the noise pattern in Fig. 2 and a lava crust temperature of 1000°C.



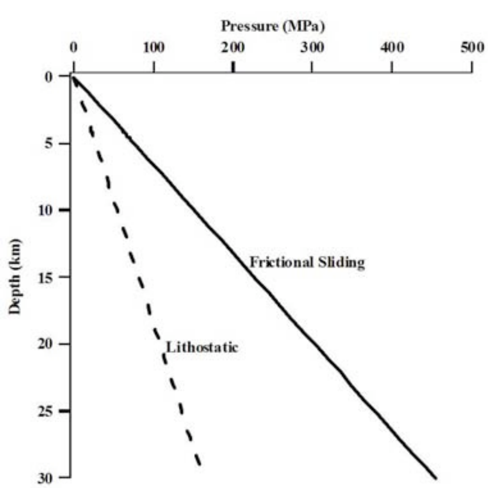
Based on this new analysis, we find no evidence for color temperatures above 1000°C at Pele. This is 200-300°C lower than previous estimates [14]. Assuming that the eruption is quite vigorous, we estimate that a surface temperature of 1000°C implies a lava temperature of ~1300°C. Curiously, this is the same result as at Tvashtar [15]. Future work will try to reconcile these lower temperatures from high-resolution observations with the higher temperatures derived from low spatial resolution data.

Superheating Mechanisms: While our improved error analysis can account for 200-300°C of the 400 °C gap between the observed temperatures and the theoretical limits, we must still account for another 100-200°C. To this end, we have searched for mechanisms by which lavas on Io can be heated as they ascend from the mantle to the surface. Previous work has rejected a number of such mechanisms [4,16]. These include (a) eruption from great depth, (b) tidal motion of the dike walls, and (c) electrical currents.

Friction from the motion of the dike walls during tidal cycles can be shown to be completely trivial [4]. Electrical currents from the ionosphere are energetic enough to heat lava at the surface, but would require odd and discontinuous geometrical arrangements of the conducting silicate melt. Furthermore, there is no observational evidence of the ionospheric currents reaching the surface of Io.

However, a combination of viscous heating during magma ascent and melt production at elevated pressure could provide significant superheating. These processes are generally negligible on the Earth, but may be important on Io because of the unique stress distribution within the Ionian lithosphere (Fig 4). The rapid resurfacing and subsidence on Io place the lithosphere under strong compression. Assuming a crust 30 km thick, ascending magma would have to overcome a confining pressure of ~0.5 GPa before being able to ascend into the crust. 20% melting at this pressure should take place at ~1270°C.

Figure 4. Stress distribution within the Ionian lithosphere. From Jaeger et al. [17].



In an ideal 1-phase system, the viscous heating of the ascending magma can be equated to the work done on the system: $Q_{visc} = \Delta P\langle v_z\rangle/L$, where Q_{visc} is the heating rate per unit time, $\langle v_z\rangle$ is the average vertical speed of the magma in the dike, and L is the vertical extent of the dike. Since rise time through the lithosphere is simply $L/\langle v_z\rangle$, the temperature increase of the magma (ΔT) can be expressed as $\Delta T = \Delta P/\rho C$, where ρ is density and C is the heat capacity. Because the magmas are expected to rise rapidly (even turbulently) through Ionian dikes [18], heat loss by conduction through the wall rocks is probably negligible and this idealized model may provide a reasonable approximation of reality.

Given a driving pressure of 0.5 GPa, magma would be heated about 100°C during ascent. If this pressure if provided by hydrostatic head, it would require a column of magma ~1000 km tall that is ~300 kg/m^3 less dense than the country rock. Overall, we predict that lavas erupting at the surface of Io will be >150°C superheated.

Conclusion: We can account for >350°C of the ~400°C discrepancy between observed and theoretically allowable lava temperatures on Io. This also removes the need for a magma ocean within Io. However, low-spatial resolution observations of Io's volcanism continue to report temperatures that are difficult to reconcile with the theoretical models. Our future work will concentrate on (a) improving our low spatial resolution temperature estimates and (b) working toward a full numerical treatment of 3-phase magma ascent in Io with coupled vent dynamics.

References:

[1] McEwen A. S. et al. (1998) *Science, 281*, 87-90. [2] Davies A. G. et al. (2001) *Icarus, 148*, 729-733. [3] Marchis F. et al. (2004) *AGU Fall Meet.* Abstract #V33C-1483. [4] Keszthelyi L. et al. (2004) *Icarus, 169*, 271-286. [5] Ross M. N. and Schubert G. (1985) *Icarus, 64*, 391-400. [6] Ojakangas G. W. and Stevenson, D. J. (1986) *Icarus, 66*, 341-358. [7] Monnereau, M. and Dubuffet F. (2002) *Icarus, 158*, 450-459. [8] Moore W. B. (2000) *Icarus, 154*, 548-550. [9] Lewis J. S. (1982) *Icarus, 50*, 103-114. [10] Consolmagno G. J. (1981) *Icarus, 47*, 36-45. [11] Keszthelyi L. and McEwen A. S. (1997) *Icarus, 130*, 437-448. [12] Ghiorso M. S. et al. (2002) *G^3, 3*, 000217. [13] McKinnon, W. B (2004) *LPS XXXV*, Abstract #2137. [14] Radebaugh et al. (2004) *Icarus, 169*, 65-79. [15] Milazzo, M. et al. (in revision) *Icarus*. [16] Kargel et al. (2003) *EOS, 84*, 313-318. [17] Jaeger W. L. et al. (2003) *LPS XXXV*, Abstract #2048. [18] Wilson L. and Head J. W. (2001) *JGR, 106*, 32997-33004.