IO'S INTERIOR: A SYNTHESIS VIEW AT THE END OF THE GALILEO ERA. L. Keszthelyi[1,2], W. L. Jaeger[1], A. S. McEwen[1], and E. P. Turtle[1], [1]Lunar and Planetary Laboratory (University of Arizona, Tucson, AZ 85721; lpk@lpl.arizona.edu), [2]*now at* U. S. Geological Survey (2255 Gemini Dr., Flagstaff AZ 86001).

Introduction: The *Galileo* spacecraft has collected its last data from Io, Jupiter's enigmatic and hyperactive moon. The new data from *Galileo*, supported by additional observations from Earth-based telescopes and the *Cassini* spacecraft, have revealed many new facets of the geologic processes acting within Io. The objective of this work is to synthesize these new data into a coherent view of the interior of Io. While we do not claim to actually know what is within Io, the picture presented here is at least consistent with the observations.

Counter-intuitively, our understanding of Io is best at the center and progressively poorer as we approach the surface. Therefore we describe Io from the inside out, ending with some hypotheses for how the processes within Io affect the surface.

Core: Our improved understanding of Io's core comes primarily from the geophysical measurements provided by *Galileo* [1, 2]. Anderson et al. [1] use the gravity data to produce two models for Io's core, one involving pure iron and the other having an Fe-FeS eutectic composition. Given the fact that sulfur is abundant on the surface of Io and that there appears to be rapid transport of material from the interior to the surface via volcanism, it seems likely that sulfur has made it to the core-mantle boundary. As such, it is highly likely that Io's core is at the Fe-FeS eutectic. In this case, the core is expected to have a density of ~5150 kg/m^3 and a radius of about 900 km.

Kivelson et al. [2] used the *Galileo* magnetometer data to show that Io has no significant intrinsic magnetic field. The lack of detectible magnetic dynamo implies that the core of Io is either completely solid or completely liquid. Given the relatively low melting temperature of the Fe-FeS eutectic composition (<1700 K at the pressure at Io's core-mantle boundary), it seems far more likely that the core is completely liquid. Thus we are quite confident that half the trip from the center of Io to the surface is through an Fe-FeS liquid.

Mantle: Anderson et al. [1] conclude that "of the major families of meteorites, only the L and LL classes have the appropriate compositions to explain the core radii and mantle densities...". If this is correct, then Io has a relatively low bulk Fe content. Given the size of the core, little iron remains in the mantle. We estimate that there would only be 5-8 wt.% FeO in Io's mantle. The resulting mineralogy would be dominated by orthopyroxene with minor olivine and clinopyroxene. This composition is not extremely different than the Earth's mantle (~8 wt.% FeO), but is less than half the iron used in earlier modeling of Io's mantle [3].

Table 1. Estimated composition of Io's mantle consistent with *Galileo* gravity data.

Oxide	L-chondrite	LL-chondrite
SiO_2	54.2	52.7
MgO	33.8	32.7
FeO*	5.0	7.8
Al_2O_3	3.1	2.9
CaO	2.5	2.5
Na_2O	1.3	1.2
TiO_2	0.2	0.2

*all iron is reported as FeO

The state of the mantle material is a matter of considerable ongoing discussion. Recent theoretical studies favor only small degrees of partial melting [4] because it is possible to model the tidal heating within a largely solid Io. However, the very high temperatures and orthopyroxene-rich compositions of Io's lavas suggested by Galileo SSI and NIMS data [5,6,7] are more consistent with large degrees of partial melting within much of Io [8].

We have conducted new model runs using the MELTS thermodynamic model [9] and the L and LL chondrite mantle compositions to quantify the degree of melting required to produce the appropriate lava compositions and temperatures. For reasons explained in the next section, we estimate that the top of the mantle is at ~1 kbar (~25 km depth). As Table 2 shows, lavas dominated by orthopyroxene only form from melts involving ~50% melting or greater. >50% melting is also needed to achieve the ≥1500 °C lava temperatures detected on Io [5,6]. However, 100% melting produces significant olivine, which is not observed in the SSI color data [7].

Table 2. Lava composition as a function of degree of partial melting of Io's mantle.

L-chondrite composition

Melting	Temp.	OPX	CPX	Ol	Plag	Other
10%	1140 °C	17%	0%	0%	61%	22%
25%	1280 °C	12%	29%	0%	55%	4%
50%	1480 °C	50%	17%	0%	29%	4%
100%	1660 °C	56%	3%	25%	16%	0%

LL-chondrite composition

Melting	Temp.	OPX	CPX	Ol	Plag	Other
10%	1142 °C	0%	20%	0%	70%	10%
25%	1270 °C	12%	28%	0%	60%	0%
50%	1470 °C	53%	12%	0%	35%	0%
100%	1657 °C	55%	0%	29%	16%	0%

Fe-FeO oxygen fugacity used.

It must be noted that there are other ways to produce the types of melts inferred to be erupted on Io. For example, if the melts were generated near the core-mantle boundary, then the appropriate compositions and temperatures could be achieved with lesser degrees of partial melting (~20 vol.%). If the lavas were rapidly erupted from such depths, they would be superheated (be hotter than the low pressure liquidus for those compositions). It is also possible that there are more exotic super-heating mechanisms on Io. The powerful electrical currents running between Io and Jupiter are capable of causing strong local heating of the lava. However, a balance of tidal heating and thermal conduction within dikes is unable to superheat the lava significantly, for reasonable dike widths and deformation rates. Instead, the added heat is mostly taken up in melting the surrounding wall rocks. It is also important to note that the detection of Mg-rich orthopyroxene (and the non-detection of olivine) in Io's lava is a tentative result based on 6-color data [7].

With these caveats, we can say that the *Galileo* data are consistent with the top of the mantle being ~50% liquid. If we assume the mantle is efficiently convecting, then the liquid fraction will decrease with depth because the melting temperature increases with pressure more quickly than the adiabatic temperature gradient. Using MELTS, and assuming the top of the mantle is 50% liquid, we estimate that the base of the mantle is at about 1600 °C and is 20% liquid. While there are many issues related to keeping this mushy magma ocean stable over geologic time, it is the model most consistent with the available observations.

Lower Lithosphere: The bulk of the lower part of the lithosphere is expected to be composed of silicates, in the form of solidified lavas and intrusive bodies. Due to the rapid resurfacing rate, heat from the mantle is unable to conduct upward faster than the lithospheric material is pushed downward. The effect is to have a very cold lithosphere with only the lowermost few kilometers undergoing significant heating [10].

The rapid burial also has the effect of horizontally compressing the lithosphere as material is pushed down into a smaller sphere [11]. This subsidence-driven compression is augmented by thermal expansion of the lowermost part of the lithosphere [12]. In order to produce the observed volume of mountains, the lithosphere must be at least 12 km thick [13]. The tectonism observed at the surface makes it exceedingly difficult for the Ionian lithosphere to be >100 km thick [13], and a thickness of 20-30 km is currently our favored estimate.

The compressive stress is sufficient to fracture rock anywhere greater than a few kilometers deep. Thus we expect most of Io's lithosphere to be broken into discrete fault-bounded blocks. From the scale of mountains, we speculate that these blocks are on the order of 100 km in lateral scale.

The compressive stress also makes it difficult for magma to ascend into and through the lithosphere. It is therefore not surprising that the new *Galileo* data show that magma appears to preferentially ascend where active tectonism has relieved some of the compressive stress [13].

Shallow Crust: Near the surface, the composition and structure of Io's crust is complicated by the addition of sulfurous volatile compounds. These compounds are capable of complex chemistry and changes between all three phases in the near-surface. The colorful and ever-changing surface of Io is a direct manifestation of the intricacy of these processes.

It is possible that the sulfurous compounds only make up a very thin veneer on the surface, but many of the new observations suggest that the volatile layer is locally at least a kilometer thick. Most striking are cliffs that appear to be "sapping" via volatilization of SO_2 at their bases but do not leave significant lag deposits. Also, Prometheus-type plumes appear to be generated by SO_2 volatilized by advancing lava flows [14]. The fact that the Prometheus plume was sustained over at least the 5-year *Galileo* Mission, during which the lavas did not move significantly onto new volatile-rich terrains, suggests that there is a substantial reservoir under the flow field.

One consequence of such a thick volatile layer is to allow a new model for patera formation. Silicate magmas would pond at the base of a low-density layer and form a sill. This sill would then heat and volatilize the overlying layer, eventually exhuming the molten sill. Once exposed, the sill would behave like a lava lake. Clow and Carr [15] showed that sulfur-rich cliffs could stand many kilometers tall as long as they were cold. The sharp vertical cliffs of some paterae may be transient features that will collapse as their bases are heated by the lava.

References: [1] Anderson J.D. *et al.* (2001) *JGR* **106**, 32963-32970. [2] Kivelson M.G. (2002) *Eos* **83**, Abstract #P22A-03. [3] Keszthelyi L. and McEwen A. (1997) *Icarus* **130**, 437-448. [4] Moore W.B. (2002) *Icarus* **154**, 548-550. [5] McEwen A. *et al*. (1998) Science 281, 87-90. [6] Davies A. *et al.* (2001) *JGR* **106**, 33079-33103. [7] Geissler P. *et al*. (1999) *Icarus* **140**, 265-281. [8] Keszthelyi L. *et al.* (1999) *Icarus* **141**, 415-419. [9] Ghiorso M.S. and Sack R.O. (1995) *Contrib. Min. Pet*. **119**, 197-212. [10] O'Reilly T.C. and Davies G.F. (1981) *GRL* **8**, 313-316. [11] Schenk P.M. and Bulmer M.H. (1998) *Science* **279**, 1514-1517. [12] McKinnon W.B. *et al.* (2001) *Geology* **29**, 103-106. [13] Jaeger W.L. *et al. JGR*, in review. [15] Kieffer S. *et al.* (2000) *Science*, **288**, 1204-1208. [16] Clow G.D. and Carr M.H. (1980) *Icarus* **44**, 268-279.