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A komatiite analog to potential ultramafic materials on Io

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Abstract. Ultramafic volcanism is one model which fits currently available data for some eruptions on Jupiter's moon, Io. Assuming that such activity does occur, it is possible to apply komatiitic analogs and terrestrial ultramafic flow models to Io. We have used such analogs and models along with Galileo data to investigate the nature of ultramafic materials that are assumed to occur at high-temperature hotspots on Io. The unusual komatiites of the Commondale greenstone belt, South Africa, are consistent with available Galileo data on the temperatures and composition of potential Ionian ultramafic materials. These komatiites had high SiO_2 (~50 wt %) and MgO (~31 wt %) contents, which are inferred to result in high liquidus temperatures (~1610°C), low dynamic viscosities (~0.2 Pa s), and low densities (~2680 kg/m^3) and in the crystallization of orthopyroxene phenocrysts (not found in other komatiites). We used the Commondale composition and the model of *Williams et al.* [1998] to investigate potential ultramafic flow behavior on Io. Compared to their terrestrial counterparts, Ionian ultramafic lava flows would have less potential for turbulent flow, lower maximum emplacement distances, and lower thermal erosion rates because of the lower Ionian gravity and their more silicic composition. Shallow (<10 m deep) thermal erosion channels are predicted to occur more proximally to the lava source than on Earth. Deep (>10 m) thermal erosion channels seem less likely on Io, unless (1) lavas are superheated, (2) lavas are more ultramafic (>31% MgO), (3) eruption rates, eruption durations, or flow volumes are high, or (4) substrates are unconsolidated, partly consolidated, or volatile-rich and thus more erodable. If volatiles are present in the lava as vesicles, then increasing lava vesicularity results in decreasing maximum emplacement distances, decreasing thermal erosion rates, and decreasing erosion channel depths for given eruption durations.

1. Introduction

The first evidence of active volcanism beyond the Earth was discovered on Jupiter's moon Io during the Voyager flybys [*Smith et al.,* 1979a,b]. Initial studies suggested that Io was dominated by low-temperature sulfur volcanism, with only minor high-temperature silicate volcanism at a few hotspots [e.g., *Nash and Howell,* 1989; *McEwen et al.,* 1989]. However, more recent telescopic and Galileo spacecraft observations [e.g., *Spencer et al.,* 1997; *McEwen et al.,* 1997; *Lopes-Gautier et al.*, 1997, 1999] show numerous high-temperature (high-*T*) hotspots, suggesting that silicate volcanism may be widespread, as proposed by *Carr* [1986]. In particular, some data indicate that there are multiple hotspots erupting materials much hotter than those found on Earth, with temperatures >1430°-1730°C [*McEwen et al.,* 1998]. These temperatures are inconsistent with historic terrestrial lavas but are consistent with either theorized superheated or prehistoric ultramafic (i.e., komatiitic) lavas [e.g., *Green,* 1975]. Thus these Io eruptions could represent the first historic occurrence of active ultramafic volcanism in the solar system, and they may provide an opportunity to study a type of volcanism that was fundamental to the Earth's early history and perhaps to that of other terrestrial planets.

In this paper we investigate the nature of Ionian lava flows, based on current understanding of their temperatures and compositions inferred from Galileo data and the fluid dynamic behavior of terrestrial analogs such as komatiites. We briefly review the evidence for high-*T* eruptions and associated deposits on Io. In particular, Galileo observations that provide temperature and compositional data that are consistent with ultramafic silicate eruptions are discussed, along with the fluid dynamic behavior of terrestrial komatiite lavas based on their physical and compositional properties. We then discuss an unusual orthopyroxene-bearing komatiite from the Commondale greenstone belt of South Africa, which we believe is the closest analog to possible Ionian ultramafic materials identified thusfar, and we discuss the composition and physical properties of this lava relative to other analogs. Finally, because available evidence supports ultramafic volcanism for some Ionian eruptions, we use the Commondale analog in an adapted version of the terrestrial ultramafic flow model of *Williams et al.* [1998] to investigate the emplacement and the potential for thermal erosion of Ionian ultramafic lavas.

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Table 1. Ionian Hotspots and Maximum Brightness Temperature Estimates

Hotspot Name	Latitude-Longitude	Maximum T^*, °C	Best Voyager Resolution, km/pixel	Inferred wt % MgO
Svarog	54°S, 270°	860	<0.5	NA
Acala	11°N, 333°	1160	<2	NA
Pele	18°S, 256°	1020	<2	NA
Pillan	10°S, 242°	>2330	2-5	>66.5
Pillan-N	9°S, 243°	1290	2-5	NA
Pillan-S	11°S, 242°	1050	2-5	NA
-	22°N, 238°	1640	2-5	32
Marduk	27°S, 209°	1340	2-5	17
Lei-kung	37°N, 203°	1350	2-5	17.5
Isum	33°N, 205°	1380	2-5	19
Amirani	23°N, 116°	>1730	5-20	>36.5
Janus	4°S, 39°	1630	5-20	31.5
Kanehekili-N	14°S, 33°	1660	5-20	33

Table modified from *McEwen et al.* [1998]. Reprinted with permission from *McEwen et al.* [1998]. Copyright 1998 American Association for the Advancement of Science. NA, not applicable.

*Maximum brightness temperatures calculated as described by *McEwen et al.* [1998].

2. Background

2.1. Evidence for High-*T* Lavas on Io

McEwen et al. [1998] summarized the evidence from Galileo observations supporting high-*T* silicate lavas on Io, including (1) solid-state imaging (SSI) eclipse and near-infrared mapping spectrometer (NIMS) images which were used to provide temperature estimates at hotspots and (2) SSI multispectral (color) images which were used to infer compositional information. The maximum brightness temperatures calculated for 13 hotspots (Table 1) show that several have temperatures much greater than typical terrestrial basaltic volcanism (<1200°C), more consistent with either superheated mafic or ultramafic volcanism [*McEwen et al.*, 1998]. Superheating of basaltic or other lavas (i.e., heating above their liquidus temperatures) is a controversial process primarily because it has not been observed on Earth historically and the geologic evidence supporting superheating is unclear. While it must remain a viable option for Io, it will not be considered further in this work. If, however, we assume that the high temperatures are indicative of magmas at or near their liquidi, we can get a rough idea of the ultramafic affinity (MgO content) of these hotspots as a function of (assumed) liquidus temperatures T_{liq} using an equation modified from *Lesher* [1983], which was modified from *Nisbet* [1982]:

$$\text{wt \% MgO} = (T_{liq} - 1000°C) / 20 \quad (1)$$

For terrestrial komatiites, equation (1) is valid for liquids >18% MgO and temperatures >1360°C, although comparisons with the program MELTS [*Ghiorso and Sack,* 1995] suggest that equation (1) gives reasonable estimates of MgO content at lower temperatures. Thus, used simply as an indicator of potential ultramafic affinity, equation (1) shows that some of the high-temperature hotspots on Io are consistent with lavas analogous to terrestrial komatiitic rocks (~12-32 wt % MgO).

In addition, Galileo SSI multispectral observations [e.g., *Geissler et al.,* 1999] show that the spectra of two hotspots (Kanehekili and Amaterasu) are similar to the spectra of enstatite, a type of orthopyroxene common in terrestrial basaltic lavas [*McEwen et al.*, 1998]. Whereas these data support the presence of silicate (rather than sulfur) volcanism on Io, they are inconsistent with typical terrestrial komatiitic volcanism because komatiites normally contain phenocryst phases of olivine and/or clinopyroxene but not orthopyroxene. Potentially, these data would indicate that the high-*T* Ionian volcanics must be quite different from terrestrial komatiites. However, as we shall see, the identification of the compositionally unusual Commondale komatiites may serve as a link between possible Ionian ultramafic volcanics and terrestrial komatiites.

2.2. Eruption Styles at High-*T* Hotspots

Voyager images of Io suggest that the products of both effusive and explosive eruptions could be identified [*Smith et al.*, 1979a,b]. Galileo images allow for the determination of surface changes at known sites of active volcanism [*Belton et al.*, 1996] and enable further study of the deposits at sites of known high-*T* eruptions. For example, Figure 1 shows a large (86 km high) plume eruption from the caldera Pillan Patera, which is one of the sites known to erupt high-temperature materials (Table 1; see also *McEwen et al.* [1998]). It is possible that this plume produced a large, 400 km diameter diffuse dark deposit around Pillan (Figure 2) that was observed in subsequent but not earlier orbits. This deposit shows some very dark material at the center, which could be lava flows, whereas the more diffuse dark material that makes up most of the deposit could be pyroclastic materials. Although the morphologies of these materials are uncertain at these low (several km/pixel) resolutions, it is clear that Galileo images of the Pillan region, known to be a site for the eruption of high-*T* materials (ranging from 1500° to >2600°C and 1280° to 1560°C on two different orbits [*McEwen et al.*, 1998]), may include both ultramafic flows and ultramafic pyroclastic deposits. Furthermore, the widespread presence of plumes on Io indicates that volatiles (e.g., SO_2) may be an important component of erupting lava compositions.

The Lei-Kung Fluctus region, in which high-*T* eruptions were also measured (1210°-1625°C [*McEwen et al.*, 1998]), may also contain examples of ultramafic flows (Figure 3). Figure 3 shows several low-albedo features covering large areas (including a >400 km long feature near the center of the image). These features are similar in appearance to long lava flows, with similarities to terrestrial flood basalts and lunar mare lavas. Once again, although we cannot be certain of either composition or morphology from this low-resolution image, ultramafic lavas are thought to have the potential to form long, widespread flows such

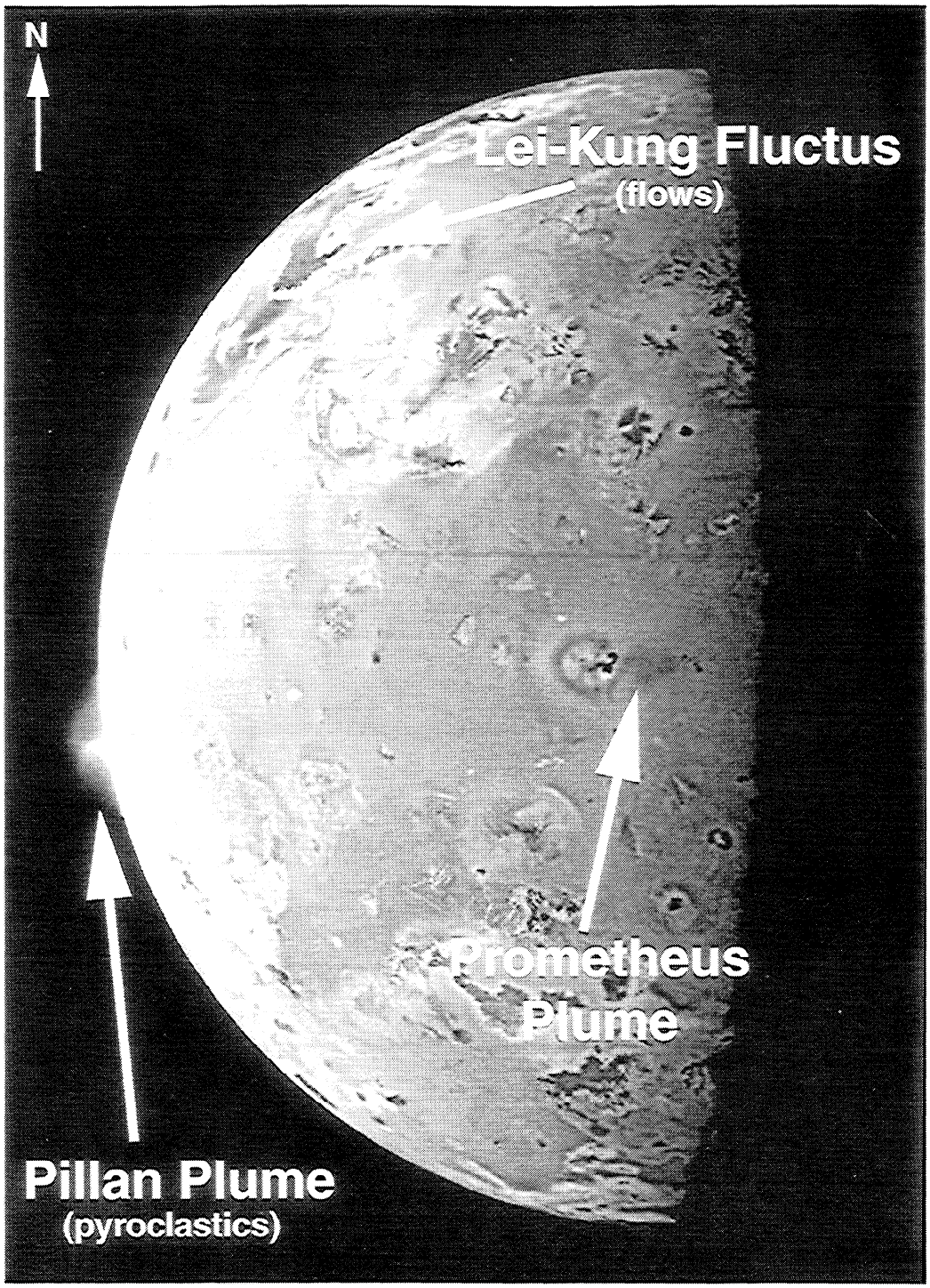

Figure 1. Galileo image of Io showing an ~86 km high eruption plume over Pillan Patera, which probably produced the dark deposit highlighted in Figure 2. Pillan is one of the sites where very high temperature eruptions were measured (ranging from 1280° to >2600°C [*McEwen et al.*, 1998]). North is to the top of the image. This image was taken during orbit C9 (June 1997) at a resolution of 2 km/pixel. This image is from Galileo press release PIA01081.

as these, because of their low-viscosity [*Huppert and Sparks*, 1985], which may have been similar to that of lunar lavas [*Williams et al.*, 1999a].

2.3. Komatiites

Komatiites are the metamorphosed remnants of ancient terrestrial ultramafic rocks, found almost exclusively in Precambrian terrains [*Arndt and Nisbet*, 1982]. Komatiites had high MgO contents (~18-32%), and possibly low dynamic viscosities (~0.1-2 Pa s), high liquidus temperatures (~1360°-1640°C), and great potential for turbulent flow and thermal erosion of underlying substrate [*Huppert et al.*, 1984; *Huppert and Sparks*, 1985; *Jarvis*, 1995; *Williams et al.*, 1998]. Although most interest in komatiites has been in understanding their

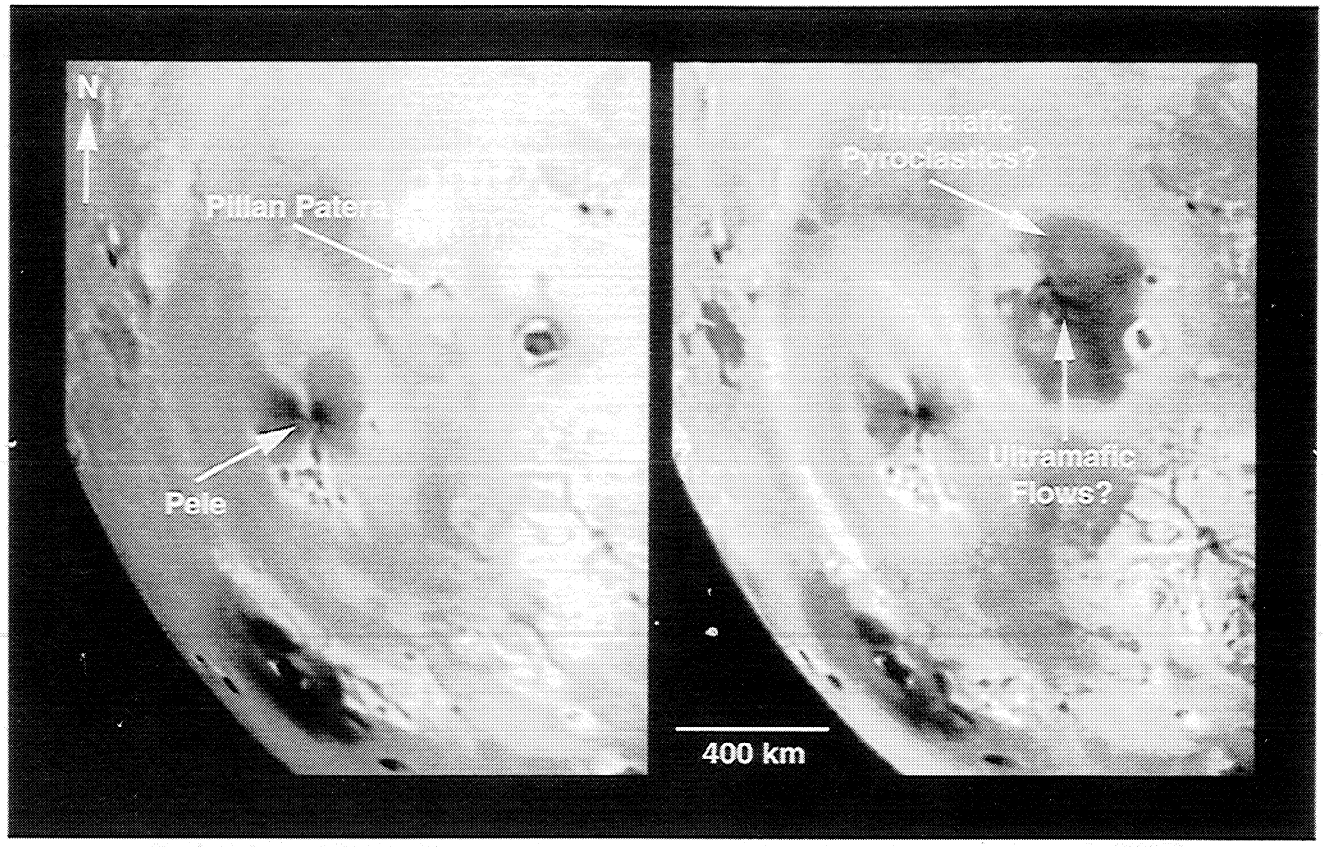

Figure 2. Galileo images of the Pele-Pillan region, taken during orbits (left) G7 (April 1997) and (right) C10 (September 1997). The large dark spot in the right image is ~400 km in diameter and may consist of ultramafic pyroclastics and lava flows from very high temperature eruptions that have occurred at Pillan. North is to the top of the images. These images were taken at resolutions of 6 km/pixel on the left and 5 km/pixel on the right. These images are from Galileo press release PIA00744.

relationship to the Archean mantle [e.g., *Nisbet et al.,* 1993; *Herzberg and O'Hara,* 1998] or to associated sulfide ore deposits [e.g., *Lesher,* 1989], komatiites are of interest to planetary geologists as analogs to low-viscosity extraterrestrial lavas that may form large lava channels by thermal erosion [e.g., *Hulme,* 1973; *Carr,* 1974; *Baker et al.,* 1992].

There is considerable variation in the thickness, extent, composition, and texture of komatiitic rocks. Komatiite lava flows range in thickness from tens of centimeters to tens of meters. Individual flows tend to have exotic petrographic textures composed of olivine and/or clinopyroxene phenocrysts, with chromite occurring as an accessory phase. Komatiites formed a wide variety of lava facies, including massive and (rare) pillowed flows, sheet flows, channelized sheet flows, lava channels, and lava ponds [*Arndt et al.,* 1979; *Lesher et al.,* 1984; *Hill et al.,* 1995], as well as subvolcanic intrusives [*Parman et al.,* 1997]. The great extent (approximately tens of meters thick and tens to hundreds of kilometers long) of some Australian komatiite sequences led to the suggestion that some komatiites may have produced large flow fields akin to the lunar maria or continental flood basalts [*Hill et al.,* 1990; 1995]. Although pyroclastic komatiitic rocks are rare, they do occur in a few areas and are associated with volatile-rich eruptions [e.g., *Gélinas et al.,* 1977; *Schaefer and Morton,* 1991].

2.4. Commondale Komatiites

Of particular interest for Io are the Commondale komatiites. The Commondale greenstone belt (~3.2-3.5 Ga [*Wilson and Carlson,* 1989; *Lopez-Martinez et al.,* 1992]) is located near the Swaziland border in South Africa. The komatiitic sequence is ~1.5 km thick and is composed of several hundred flow units. Most of the outcrops are of poor quality, so that identification of textural zones is done mostly from drill core. The morphology of individual komatiite flow units includes (from top to bottom) a thin chilled flow top, a spinifex-textured zone consisting of exclusively orthopyroxene or orthopyroxene with platy olivine crystals, a mixed olivine cumulate-textured zone (which also contains 2-10% orthopyroxene spinifex crystals) that grades downward into cumulus olivine crystals, a basal olivine cumulate with splays of orthopyroxene crystals, and a lower chill zone. Individual flow units range from 1 to 13 m thick, although most flows are typically 1-3 m thick. The orthopyroxene spinifex crystals are characteristic of the whole sequence, which appears to be the only komatiite locality containing orthopyroxene phenocrysts. The sampled olivines have compositions up to Fo_{93} (i.e., 93% Mg-rich olivine), and the orthopyroxenes have compositions up to En_{94} (i.e., 94% Mg-rich orthopyroxene), indicative of crystallization from very Mg rich liquids.

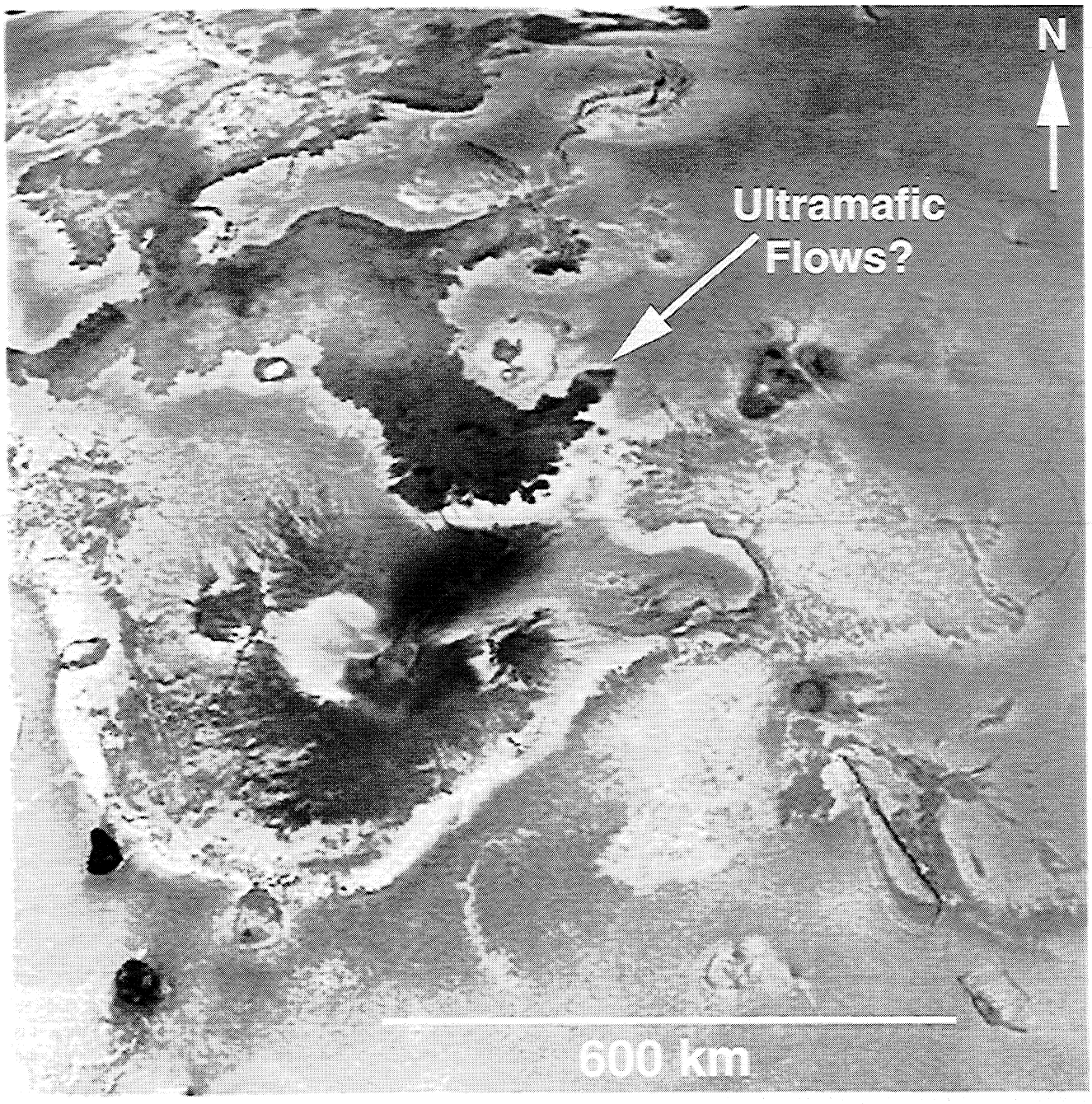

Figure 3. Galileo image of the Lei-Kung Fluctus region, taken during orbit C3 (November 1996). The dark flow near center is over 400 km long and may have been produced by a high-temperature, possibly ultramafic eruption like the one measured later during orbit E11 (i.e., temperatures within the range 1210°-1625°C [*McEwen et al.*, 1998]). North is to the top of the image. This image was taken at a resolution of 2.5 km/pixel, and is from Galileo press release PIA00537.

3. Methodology

3.1. Calculation of Physical Properties

On the basis of petrologic analyses of samples collected at Commondale, the parental lava composition has been estimated (Table 2). Using algorithms from experimental petrology, it is possible to calculate the thermal and rheological properties of the Commondale komatiites relative to other lavas. As summarized by *Williams et al.* [1998], the lava dynamic viscosity is calculated using the algorithm of *Shaw* [1972]. The liquidus temperature is calculated using MELTS [*Ghiorso and Sack*, 1995], and the solidus temperature for komatiite is from *Arndt* [1976]. We calculate the lava density using the method of *Bottinga and Weill* [1970] with the partial molar volume data of *Mo et al.* [1982], the specific heat is calculated from the heat capacity data of *Lange and Navrotsky* [1992], and the heat of fusion is estimated using

Table 2. Inferred Liquid Compositions and Physical Properties for Several Komatiitic and Basaltic Lavas

	Commondale Komatiite	Kambalda Komatiite	Barberton Komatiite	Cape Smith Kom. Basalt	Lunar Mare Basalt	Tholeiitic Basalt
Components						
SiO_2	49.8	45.0	47.9	46.9	43.6	50.9
TiO_2	0.1	0.3	0.4	0.6	2.6	1.7
Al_2O_3	7.9	5.6	4.1	9.8	7.9	14.6
Fe_2O_3	0.5	1.4	1.9	-	-	-
FeO	4.8	9.2	9.7	14.4	21.7	14.6
MnO	0.1	0.2	0.2	0.3	0.3	-
MgO	30.9	32.0	27.5	18.9	14.9	4.8
CaO	5.2	5.3	7.5	8.6	8.3	8.7
Na_2O	0.4	0.6	0.2	0.3	0.2	3.1
K_2O	0.01	0.03	0.02	0.05	0.05	0.8
Physical Properties						
T_{liq}, °C	1611	1638	1556	1419	1440	1160
T_{sol}, °C	1170	1170	1170	1150	1150	1080
ρ at T_{liq}, kg/m^3	2680	2770	2760	2800	2900	2750
c, J/kg·°C	1780	1790	1740	1640	1570	1480
μ at T_{liq}, Pa s	0.22	0.08	0.20	0.74	0.40	86
L at T_{liq}, J/kg	6.84E+05	6.97E+05	6.58E+05	5.96E+05	6.06E+05	5.37E+05
k at T_{liq}, J/m s °C	0.4	0.4	0.4	0.5	0.5	0.7
h, m	10	10	10	10	10	10
u at T_{liq}, m/s	4.6	4.8	4.5	3.9	4.2*	1.7
Re at T_{liq}	5.74E+05	1.74E+06	6.31E+05	1.46E+05	3.09E+05*	5.50E+02
Pr at T_{liq}	1.0E+03	3.8E+02	8.4E+02	2.6E+03	1.3E+03*	1.9E+05
h_T at T_{liq}, J/m^2 s °C	260	396	277	172	217*	N/a
u_{mb} at T_{liq}, m/day	1.1	1.7	1.0	0.52	0.24*	N/a
Composition location	Commondale, South Africa	Kambalda, Western Australia	Barberton, South Africa	Katinniq, Cape Smith Belt, Canada	Apollo 12 sample 12002	Columbia River Basalt, Washington
Composition reference	this study	*Lesher and Arndt* [1995]	*Viljoen et al.* [1983]	*Barnes et al.* [1982]	*Walker et al.* [1976]	*Murase and McBirney* [1973]

T_{liq}, liquidus temperature; T_{sol}, solidus temperature; ρ, density; *c*, specific heat; μ, dynamic viscosity; *L*, heat of fusion; *k*, thermal conductivity; *h*, flow thickness; *u*, flow velocity; *Re*, Reynolds number; *Pr*, Prandtl number; h_T, convective heat transfer coefficient; u_{mb}, thermal erosion rate of consolidated basalt, NA, not applicable for nonturbulent flows. Read 6.84E+05 as 6.84 x 10^5.
*Erupted on Earth.

data from *Navrotsky* [1995]. The thermal conductivity is calculated using a curve fit to data from *Snyder et al.* [1994] and *Murase and McBirney* [1973]. From these properties it is possible to estimate (for a given flow thickness) the flow velocity, Reynolds number, Prandtl number, and convective heat transfer coefficient (see section 3.2 and *Williams et al.* [1998]) and compare these properties to those of other komatiites. An estimate of the maximum thermal erosion rate of basaltic substrate can also be made.

The typical range of terrestrial low-SiO_2, high-MgO komatiites is represented by the Barberton and Kambalda compositions (Table 2). In comparison, the Commondale komatiites have a higher SiO_2 content (~50% compared to 45-48%) and an intermediate MgO content (~31% compared to 28-32%) relative to these komatiites, and this unique composition has specific effects on the thermal and rheological properties of the lava. For example, the liquidus temperature for the Commondale komatiites (~1610°C) is ~50°C hotter than the Barberton komatiites, although it is significantly cooler than that of the Kambalda komatiites (~1640°C), which may have been the highest-MgO liquids erupted on Earth. The higher SiO_2 and lower FeO contents of the Commondale komatiites result in a lower liquid density than other komatiites, and presumably the higher SiO_2, higher MgO, and lower FeO contents are also responsible for the production of orthopyroxene as a primary crystallization phase. On the basis of the flow thicknesses observed in drill core the Reynolds numbers for the Commondale komatiites suggest that the thicker (e.g., 3-10 m) flows should have erupted turbulently, similar to most other komatiite flows. However, the convective heat transfer coefficient is somewhat lower than that of other komatiites, because of the atypical composition of the Commondale komatiites. This indicates that their thermal erosion potential should have been somewhat less than that of other komatiites and that these lavas were probably less capable of forming thermal erosion channels in basaltic substrate. We will investigate this further in section 4. Using the compositions for the Commondale and Kambalda komatiites, it is possible to model the emplacement of these lavas on Io and Earth.

3.2. Modeling

We have adapted the model of *Williams et al.* [1998] to investigate the emplacement and thermal erosion potential of inferred ultramafic lavas on Io. Given a set of initial conditions (i.e., lava and substrate major oxide composition; lava eruption, liquidus, and solidus temperatures; and lava flow thickness, ground slope, ground melting temperature, and ambient temperature), our model calculates the initial (at vent) values of important temperature- and composition-dependent thermal and rheological properties of the lava and substrate (i.e., density, viscosity, specific heat, thermal conductivity, and heat of fusion) using the algorithms described in section 3.1 (see also Table 2). A series of auxiliary equations are then used to calculate additional lava properties, including crystallinity *X*

$$X = \frac{T_{liq} - T}{T_{liq} - T_{sol}}; \quad (2)$$

bulk viscosity μ_b

$$\mu_b = \mu_l\left(1-\frac{X}{0.6}\right)^{-2.5} \qquad X<0.3 \tag{3}$$

$$\mu_b = \mu_l \exp\left\{\left[2.5+\left(\frac{X}{0.6-X}\right)^{0.48}\right]\frac{X}{0.6}\right\} \qquad X\geq 0.3 \tag{4}$$

[*Pinkerton and Stevenson*, 1992]; flow velocity u, friction coefficient λ, and Reynolds number Re (solved iteratively)

$$u=\sqrt{\frac{4gh\sin(\psi)}{\lambda}} \tag{5}$$

$$\lambda=\left[0.79\ln(Re)-1.64\right]^{-2} \tag{6}$$

$$Re=\frac{\rho_l hu}{\mu_b}; \tag{7}$$

Prandtl number Pr

$$Pr=\frac{c_l\mu_b}{k_l}; \tag{8}$$

convective heat transfer coefficient h_T

$$h_T=\frac{0.027k_{\text{eff}}\,\text{Re}^{4/5}\,\text{Pr}^{1/3}}{h}\left(\frac{\mu_b}{\mu_g}\right)^{0.14}; \tag{9}$$

lava erosion rate u_m

$$u_m=\frac{h_T\left(T-T_{mg}\right)}{E_{mg}}; \tag{10}$$

and degree of contamination of the lava by assimilated substrate $S(x)$

$$S(x)=1-\frac{Q_0}{Q(x)},\quad Q(x)=Q_0+\int_0^x u_m dx. \tag{11}$$

Flow rate is conserved, such that decreasing flow velocity results in increasing flow thickness. Material added by assimilation also acts to increase thickness while decreasing velocity.

The lava density, specific heat, flow thickness, flow velocity, and heat transfer coefficient are used to determine the heat loss from the model flow by numerically solving a first-order ordinary differential equation giving lava temperature as a function of distance downstream [*Huppert and Sparks*, 1985]:

$$\rho_b c_l hu\frac{dT}{dx}=-h_T\left(T-T_{mg}\right)-h_T\left(T-T_{\text{sol}}\right)-\frac{\rho_b c_l h_T\left(T-T_{mg}\right)^2}{E_{mg}} +\rho_b c_l hu\frac{dT}{dx}\frac{L_l x'(T)}{c_l} \tag{12}$$

$$E_{mg}=\rho_g\left[c_g\left(T_{mg}-T_a\right)+L_g\right]. \tag{13}$$

Equation (12) assumes that heat transfer occurs (1) by losses due to convection within the turbulent lava to the top and base of the flow and by thermal erosion (ablation) that removes the underlying substrate and (2) by gains due to release of latent heat of crystallization. At each model distance increment the compositional change in the liquid lava due to the assimilation of thermally eroded substrate (S) can be determined using the mass balance expression

$$M_{\text{new}}=M_{\text{old}}(1-\Delta S)+M_{\text{asm}}(\Delta S), \tag{14}$$

and the compositional change in the liquid lava due to the crystallization of olivine and/or orthopyroxene (X) can be determined using the mass balance expression

$$M_{\text{new}}=M_{\text{old}}(1-\Delta X)-\left[p_{\text{olv}}M_{\text{olv}}+p_{\text{opx}}M_{\text{opx}}\right](\Delta X) \tag{15}$$

in conjunction with partition coefficient and stoichiometric algorithms. Olivine-liquid and orthopyroxene-liquid partition coefficients for Fe^{2+}, Mg, and Ca are from *Beattie et al.* [1991, 1993]; coefficients for Ti and Al are from *Kennedy et al.* [1993]. Equations (14) and (15) result in a new liquid lava composition that is used to recalculate the temperature- and composition-dependent thermal, rheological, and fluid dynamic properties of the lava at each model increment. Thus the physical and geochemical evolution of the lava flow with distance is simulated.

On Earth, because field evidence suggests subaqueous emplacement of komatiitic flows at most localities [e.g., *Lesher*, 1989], crustal growth is constrained by convective cooling by overlying seawater off of the growing upper surface crust. On Io, crustal growth is constrained by assuming radiative cooling to a vacuum, in which the rate of crustal growth is given by

$$\frac{dh_c}{dt}=u\frac{dh_c}{dx}=\frac{\left[\varepsilon\sigma\left(T_c^4-T_a^4\right)\right]-h_T\left(T-T_{\text{sol}}\right)}{\rho_b\left[c_l\left(T-T_{\text{sol}}\right)+L_l\right]}. \tag{16}$$

Now at some distance from the vent a balance is achieved between the heat fluxes in (16). At that point, $dh_c/dt = 0$ and the steady state crustal temperature T_{cs} can be found by equating the convective heat flux in the lava with the radiative heat flux to space

$$\varepsilon\sigma\left(T_{cs}^4-T_a^4\right)=h_T\left(T-T_{\text{sol}}\right). \tag{17a}$$

Solving for the steady state crustal temperature T_{cs} gives

$$T_{cs}=\sqrt[4]{T_a^4+\frac{h_T\left(T-T_{\text{sol}}\right)}{\varepsilon\sigma}}. \tag{17b}$$

Having calculated the steady state crustal temperature, the steady state crustal thickness h_{cs} is found to be

$$h_{cs}=\frac{k_c\left(T_{\text{sol}}-T_{cs}\right)}{\varepsilon\sigma\left(T_{cs}^4-T_a^4\right)}. \tag{18}$$

This calculation allows us to estimate roughly the thickness of the crust on the flow at any given distance from the eruption source, based on the heat loss from the turbulent lava into the cold Ionian environment.

Because of the widespread presence of potentially volatile-rich plumes and the potential for extreme vesiculation during lava emplacement in a vacuum (*A. Harris*, personal communication,

Table 3. Input Values for Modeling Ultramafic Lava Emplacement on Io and Earth

Parameter	Io	Earth
Lava composition	Commondale komatiite	Kambalda komatiite
Eruption temperature, °C	1611	1638
Liquidus temperature, °C	1611	1638
Solidus temperature, °C	1170	1170
Flow thickness, m	10	10
Emplacement environment	vacuum	submarine
Heat loss off of upper flow surface	radiation	convection
Crystallizing phases	90% olivine, 10% Opx	100% olivine
Substrate composition	consolidated basalt	consolidated basalt
Substrate slope	0.1°	0.1°
Gravity, m/s^2	1.797	9.812

Opx, orthopyroxene. See Table 2 for lava compositions.

1999) it is appropriate to investigate the effect of vesicles on ultramafic flow behavior on Io. The presence of bubbles in lava should increase viscosity at low strain rates [*Pinkerton and Stevenson*, 1992] and decrease lava density and specific heat. By defining a parameter for the fraction of the lava consisting of vesicles f_v the effect on these physical properties can be assessed by the following equations:

$$\rho_{\text{eff}} = f_v(\rho_{\text{gas}}) + (1 - f_v)\rho_b \tag{19}$$

$$c_{\text{eff}} = f_v(c_{\text{gas}}) + (1 - f_v)c_l \tag{20}$$

$$\mu_{\text{eff}} = \mu_b \left[\frac{1}{1 - (1.3 f_v)^{1/3}} \right], \tag{21}$$

in which the subscript "eff" refers to liquid plus crystals and/or vesicles, "gas" refers to the volatile gas in the vesicles, "b" refers to bulk (liquid + crystals), and "l" refers to liquid. Equation (21), from *Sibree* [1934], is valid for foams with values of f_v up to 75%, and *Pinkerton and Stevenson* [1992] suggest that it is still a useful algorithm to assess the effect of bubbles on lava viscosity. Because SO_2 is the most abundant volatile gas on Io, it is the logical choice for our model volatile component. Although very high temperature measurements of the thermal and rheological properties of SO_2 do not exist, the effect of this volatile gas on lava emplacement can still be assessed with available data: ρ_{gas} = 2.927 kg/m^3 and c_{gas} = 886 J/(kg K) at 1500 K [*Lide*, 1993].

We have run models using the Commondale and Kambalda compositions to investigate the effect of compositional differences on ultramafic flow behavior on Io (Table 3). We have also run models using these compositions under terrestrial emplacement conditions to identify any planetary/environmental effects on lava emplacement. Finally, we have run models for various degrees of vesicularity (0%, 30%, and 70%) in our ultramafic composition to assess the effect of vesicles on lava emplacement.

4. Results

Our modeling results (Figure 4 and Table 4) show that a Commondale-type lava should erupt as a turbulent flow on Io. For example, an initially 10 m thick flow should be capable of traveling turbulently almost 600 km, assuming a sufficiently flat and unobstructed basaltic surface. This distance is less than that predicted for a Kambalda-type komatiite erupted under the same conditions (~720 km) and is much less than that predicted for these lavas erupted on Earth (~930 km for Commondale and >1000 km for Kambalda). These results are due to several factors. First, the higher SiO_2, lower MgO content of the Commondale versus Kambalda komatiites produces a lower liquidus (1610°C versus 1640°C) and higher viscosity (0.2 Pa s versus 0.08 Pa s), which results in a reduced potential for turbulent flow and thermal erosion. Second, the lower Ionian gravity relative to Earth (1.8 m/s^2 versus 9.8 m/s^2) causes flows of the same thickness to erupt at lower flow rates and have slower flow velocities (e.g., ~2.2 m/s or 8 km/h on Io versus ~4.4 m/s or 16 km/h on Earth). Thus a Commondale-type komatiite would generate on Io (relative to Earth) a lower Reynolds number (~280,000 versus ~540,000), a lower maximum erosion rate (0.6 m/d versus 1.1 m/d), and a slightly lower degree of lava contamination (5% versus 6%) by melted and assimilated substrate. To form a thermal erosion channel of a given depth on Io, ultramafic lavas must therefore flow for longer durations than their terrestrial counterparts, and deep erosion channels should be restricted to locations more proximal to the lava source than their terrestrial counterparts. For example, with a maximum thermal erosion rate (at vent) of ~60 cm/d for flow over a consolidated, basaltic substrate, and assuming sustained flow for a week, the Commondale-type lava on Io would theoretically be capable of producing an erosion channel ~3-4 m deep within ~40 km of the vent. In comparison, a terrestrially emplaced Commondale lava could have (theoretically) attained a maximum turbulent flow distance of >900 km and, with a maximum thermal erosion rate of 110 cm/d and sustained flow for a week, could have formed an erosion channel ~3-4 m deep within ~85-140 km of the vent.

If a Commondale-type lava erupted on Io with a significant fraction of vesicles, then our model results (Figure 5) predict that for increasing degrees of vesicularity, lava density will decrease, specific heat will decrease, and viscosity will increase, resulting in lower Reynolds numbers, higher Prandtl numbers, and lower heat transfer coefficients. This has the effect (for a lava flow with 30% vesicles relative to a nonvesiculated flow) of producing lower maximum turbulent flow distances (390 versus 590 km), lower maximum thermal erosion rates (0.3 versus 0.6 m/d), lower erosion channels depths at a given time (after one week, 2.0 versus 4.3 m), and lower degrees of lava contamination by substrate (2.4 versus 4.9%). A heavily vesiculated flow (70% vesicles) would have even lower maximum turbulent flow distances (~30 km), maximum thermal erosion rates (~0.06 m/d), erosion channel depths (after one week, ~0.5 m), and degrees of lava contamination by substrate (0.1%).

5. Discussion

Our results (Figure 4) show that the maximum turbulent emplacement distances, erosion rates, and erosion depths for Ionian ultramafic flows would be measurably less than for

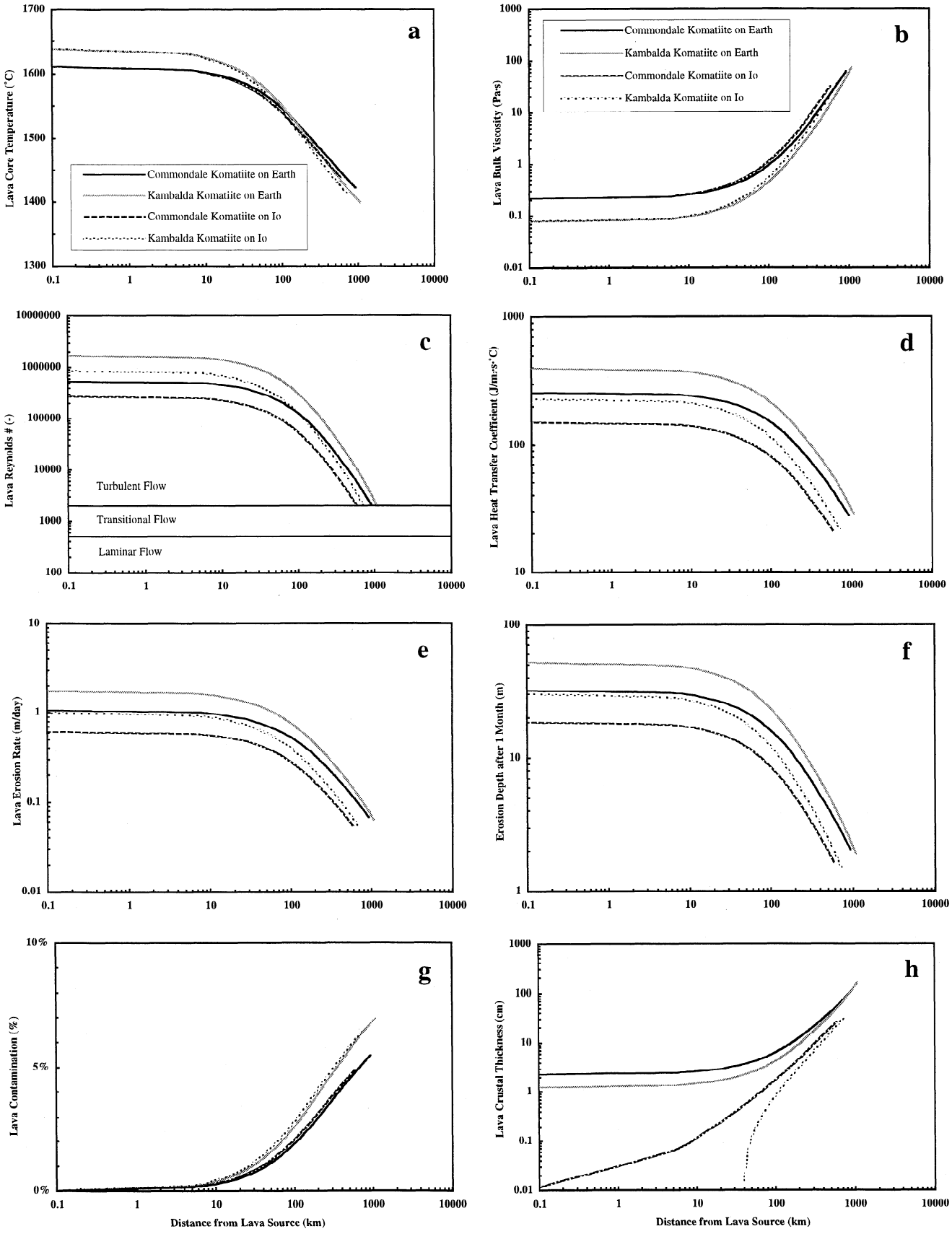

Figure 4. Model results for the emplacement of initially 10 m thick komatiite lava flows over basaltic substrate on Io and Earth. Ground slopes are assumed to be 0.1°. Terrestrial komatiites were emplaced under water; Ionian komatiites were emplaced in a vacuum.

Table 4. Model Results for the Emplacement of Commondale- and Kambalda-Type Komatiite Lava Flows Over Basaltic Substrate on Earth and Io

Property	Commondale Komatiite on Earth	Kambalda Komatiite on Earth	Commondale Komatiite on Io	Kambalda Komatiite on Io
Maximum flow distance,* km	930	1080	590	720
Maximum flow thickness,* m	17	18	16	17
Maximum erosion rate,† m/day	1.1	1.7	0.6	1.0
Maximum erosion depth after one week,† m	7.4	12	4.3	7.0
Maximum contamination,* %	5.5	7.0	5.0	6.4
Maximum crustal thickness,* cm	120	160	26	31

*As turbulent flow, limit at Reynolds number of 2000.
†At vent.

terrestrial komatiite flows of the same composition and thickness. These results are due to the lower gravity of Io relative to Earth, which results in lower flow rates, lower Reynolds numbers, and lower convective heat transfer coefficients and thus in lower thermal erosion rates on Io relative to Earth, all else being equal. Because the Commondale komatiites have a lower emplacement and erosional potential than Kambalda komatiites on Earth as well, due to their lower ultramafic affinity (i.e., lower MgO content (31 versus 32% MgO) and higher silica content (~50 versus 45% SiO_2)) and thinner flows (1-13 m thick), it seems unlikely that deep thermal erosion channels in basalt would have been produced on Earth. This is consistent with the available field and drill core data from Commondale that show no evidence of lava channels. Recent studies [*Barnes et al.*, 1988; *Williams et al.*, 1999b] suggest that either large-flow-rate or large-volume eruptions (i.e., with flows >10-100 m thick) are required to produce deep (tens to hundreds of meters) channels in basaltic substrate, and that thinner flows (≤10 m thick) are unlikely to be capable of such large-scale thermal erosion. Most evidence from Kambalda [e.g., *Groves et al.*, 1986; *Frost and Groves*, 1989; *Williams et al.*, 1998] suggests that komatiites thermomechanically eroded a few meters of unconsolidated, water-saturated sediment in preexisting lava channels or topographic depressions on the basaltic ocean floor, rather than eroding deep channels in basalt.

Applying this to Io, if the Commondale komatiites are typical of Ionian ultramafic lavas, we predict that Ionian ultramafic flows, assuming that they are not superheated, should be emplaced as short turbulent flows, with maximum turbulent flow distances of less than a few hundred kilometers. Thermal erosion channels, if they form, will occur proximal to the eruption source (e.g., within 100 km). As *Williams et al.* [1998, 1999b] show, larger, deeper erosion channels are possible if (1) flow rates are high (e.g., >500 m^3/s), (2) flow durations are very long (weeks to months), or (3) substrates are unconsolidated or volatile-rich such that mechanical erosion becomes important [*Williams et al.*, 1998; *Cas and Self*, 1998]. Because some hotspots have been active for years to decades [*McEwen et al.*, 2000] and individual events like the Pillan eruption of 1997 may have lasted at least several weeks, we cannot rule out the possibility that long (>100 km) thermal erosion channels are possible.

If volatiles like SO_2 are present in Ionian ultramafic lavas as vesicles, then we predict that even shorter, less turbulent flows should be emplaced, probably without forming deep thermal erosion channels. The changes in the thermal and rheological parameters of the lava due to increasing vesicularity (increasing viscosity and decreasing density and specific heat) act to decrease the turbulent emplacement and thermal erosion potential of ultramafic lavas. For very high vesicularities (>75%), erupting lavas could have such high viscosities that only laminar emplacement may be possible. These results do not consider the potential cooling effects of vesicles; *Keszthelyi* [1994] predicted from modeling and field studies of vesiculated Hawaiian pahoehoe flows that thermal radiation across vesicles should have significant cooling effects at high temperatures (T >800°C). If these results apply to Io, even shorter vesicular ultramafic flows than modeled here could be produced. All of this assumes that the Ionian ultramafic flows retain high vesicle contents during emplacement. On Earth, although some komatiite flows have vesicular zones in their interiors [*Beresford et al.*, 1998, 1999], many komatiite flows have vesicular zones restricted to the very tops of flows. Thus degassing may occur very rapidly and may not markedly effect the rheological properties of the flows during emplacement. However, the eruption plume at Pillan Patera (Figure 1) and the dark Pillan deposit (Figure 2) suggest that there could be areas of substantial explosive ultramafic volcanism on Io. True komatiitic pyroclastic eruptions were rare in the Earth's Archean, and they appear to be associated with unusual, volatile-rich eruptions. On Earth, H_2O [*Gélinas et al.*, 1977] or CO_2 [*Schaefer and Morton*, 1991] appear to have been the volatile components in the eruptions. On Io, S or SO_2 might play that role (S.A. Fagents et al., Implications of ultramafic compositions for explosive volcanism on Io, submitted to *Eos, Trans. Am. Geophys. Un.*, Fall Meeting Abstracts, 1999).

Finally, the widespread presence of sulfur compounds on the surface of Io brings up another interesting aspect of potential ultramafic volcanism. Their presence begs the question of whether komatiite-associated sulfide ores could form. In the Earth's Archean, komatiite-hosted sulfides formed by thermal erosion of sulfur-rich substrates by metal-rich komatiites, followed by the segregation of immiscible sulfides and miscible silicate components, the scavenging of metals from komatiites, and the final accumulation of dense metal sulfide melts at the base of the komatiite flows in lava channels or other topographic depressions (see reviews by *Lesher* [1989] and *Lesher and Campbell* [1993]). Although there have been no sulfide ore deposits discovered within the Commondale komatiites, the strong depletion of NiO in the olivines of some flows suggests that sulfides may have formed sometime during the emplacement of these lavas. Thus, although the elemental geochemistry and thermodynamics of any Ionian ultramafic materials are unknown, it may be possible that such deposits could form in metal-bearing Ionian lavas.

On the basis of current information a high-temperature, orthopyroxene-bearing ultramafic composition similar to Commondale continues to be the best interpretation of Galileo

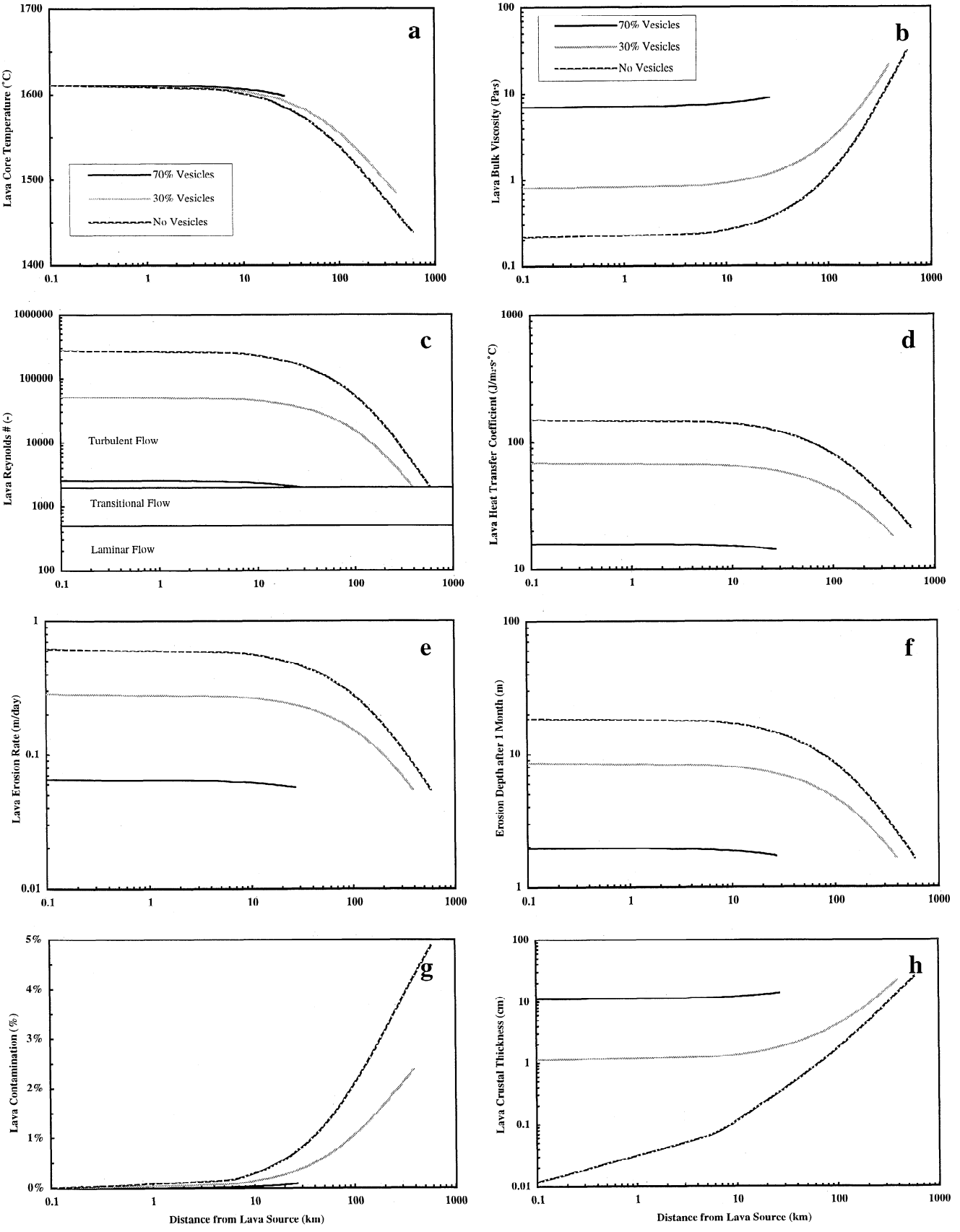

Figure 5. Model results for the emplacement of initially 10 m thick komatiite lava flows with various vesicularities over basaltic substrate on Io. Ground slopes are assumed to be 0.1°.

multispectral data for some very hot Ionian lavas. However, it is possible that this composition resulted from contamination of a very high MgO (>32 wt %) melt by Ionian crust or mantle material, and it is possible that other, perhaps more ultramafic compositions exist at Ionian hotspots. Additional information on the morphologies and compositions of Ionian ultramafic flows from anticipated Galileo encounters may help to constrain further the nature of Ionian high-T volcanism.

6. Summary

Using terrestrial analogs and Galileo data on the nature of high-T lavas, we have investigated the physical properties and emplacement style of potential ultramafic lavas on Io. The closest analog to nonsuperheated, high-T Ionian materials with enstatite-like spectral signatures [*McEwen et al.*, 1998] is an unusual orthopyroxene-bearing komatiite from the Commondale greenstone belt of South Africa. These lavas had high SiO_2 contents (~50 wt %) and high MgO contents (~31 wt %), resulting in inferred high liquidus temperatures (~1610°C), low dynamic viscosities (~0.2 Pa s), low densities (~2680 kg/m^3), and the crystallization of orthopyroxene phenocrysts, which have not been observed in other komatiites. We adapted the model of *Williams et al.* [1998] to simulate emplacement on Io's surface and used the Commondale composition to investigate potential ultramafic flow behavior on Io and Earth. Because of the lower Ionian gravity relative to Earth and because of the more silicic composition of the Commondale lavas relative to other high-MgO komatiites (e.g., Kambalda), Ionian ultramafic lava flows would be less turbulent and have lower maximum emplacement distances and lower thermal erosion rates relative to their terrestrial counterparts. Shallow (<10 m) thermal erosion channels would probably occur more proximally to the lava source than their terrestrial counterparts, and deep (>10 m) thermal erosion channels would probably occur only if (1) the lavas are superheated and thus capable of flowing longer distances at higher erosion rates; (2) the lavas are more ultramafic (>31% MgO); (3) the lavas were erupted at high effusion rates, with high flow volumes, or with long flow durations; or (4) the lavas thermally eroded unconsolidated, partly consolidated, or volatile-rich substrates, which are easier to thermally (and/or mechanically) erode than consolidated substrates [*Williams et al.*, 1998, 1999b]. Conditions 3 and 4 above have been noted on Io (L. Keszthelyi, personal communication, 1999). If Ionian ultramafic lavas were vesicle-rich, resulting in higher-viscosity, lower density flows, then even shorter, less turbulent, less erosive, perhaps unchannelized flows would be predicted on Io.

Notation

c_l	lava specific heat, J/kg °C.
c_g	substrate specific heat, J/kg °C.
E_{mg}	energy required to melt substrate, J/m^3.
ε	lava emissivity.
f_v	fraction of vesicles in lava.
g	gravitational acceleration, m/s^2.
h	lava flow thickness, m.
h_c	lava crustal thickness, cm.
h_{cs}	lava steady state crustal thickness, cm.
h_T	lava convective heat transfer coefficient, J/(m^2 s °C).
k_c	lava crust thermal conductivity, J/(m s °C).
k_{eff}	lava effective thermal conductivity, J/(m s °C).
k_l	lava thermal conductivity, J/(m s °C).
L_g	substrate heat of fusion, J/kg.
L_l	lava heat of fusion/crystallization, J/kg.
λ	lava friction coefficient.
M_{asm}	substrate composition.
M_{new}	lava composition in current model increment.
M_{old}	lava composition in previous model increment.
M_{olv}	olivine composition in current model increment.
M_{opx}	orthopyroxene composition in current model increment.
p_{olv}	proportion of olivine in model lava.
p_{opx}	proportion of orthopyroxene in model lava.
μ_b	lava bulk viscosity, Pa s.
μ_g	substrate melt viscosity, Pa s.
μ_l	lava liquid viscosity, Pa s.
ψ	slope of the ground, degrees.
Pr	lava Prandtl number.
ρ_b	lava bulk density, kg/ m^3.
ρ_g	substrate density, kg/ m^3.
ρ_l	lava density, kg/ m^3.
Q_0	initial flow rate, m^2/s.
$Q(x)$	flow rate, m^2/s.
Re	lava Reynolds number.
σ	Stefan-Boltzmann radiative constant, J/(m^2 s °C^4).
$S(x)$	degree of lava contamination by substrate.
t	time since flow began, s.
T	lava temperature, °C.
T_a	ambient temperature of the environment, °C.
T_c	lava upper surface crustal temperature, °C.
T_{cs}	lava steady state crustal temperature, °C.
T_{liq}	lava liquidus temperature, °C.
T_{mg}	substrate melting temperature, °C.
T_{sol}	lava solidus temperature, °C.
u	lava flow velocity, m/s.
u_m	erosion rate of the substrate, m/s.
x	distance from source vent, m.
X	volume fraction of crystals.
$X'(T)$	rate of change of crystal fraction with temperature.

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