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A reassessment of the emplacement and erosional potential of turbulent, low-viscosity lavas on the Moon

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Abstract. We have reevaluated the role of thermal erosion by low-viscosity lunar lavas as a mechanism for the formation of the lunar sinuous rilles. We have adapted the model of *Williams et al.* [1998] and used the compositions of an Apollo 12 basalt and a terrestrial komatiitic basalt to investigate the compositional and environmental effects on the flow behavior of low-viscosity lavas on the Moon and the Earth. Our model predicts that lunar lavas could have erupted as turbulent flows that were capable of flowing hundreds of kilometers on a sufficiently flat, unobstructed substrate. These results are consistent with previous studies. Modeling of lavas over a substrate of the same composition shows that thermal erosion rates would have been low (~10 cm d^{-1}). As a result, long-duration eruptions (approximately months to years) would have been required to incise deep (tens to hundreds of meters) channels. Partial melting and mechanical removal of the substrate, a mechanism suggested by *Hulme* [1973] to enhance erosion, only slightly increases thermal erosion rates. Other factors, such as higher flow rates or lava superheating, could have produced deep rilles by thermal erosion during shorter-duration eruptions. A superheated lunar lava not only would have had a higher erosion rate (~40 cm d^{-1}) but also would have remained uncrusted for tens of kilometers, which is consistent with the open channel morphology of most sinuous rilles. For lunar lavas with large volatile (i.e., vesicle) contents, the presence of vesicles would have tended to increase viscosity at low strain rates, resulting in shorter turbulent flow distances, lower thermal erosion rates, and thus shallower erosion channel depths for given eruption durations.

1. Introduction

Some of the most enigmatic features found on the Moon are the lunar sinuous rilles. First observed in the eighteenth century, these unusual, channel-like features were studied in the late nineteenth and early twentieth centuries using high-quality telescopic images [e.g., *Neison*, 1876; *Pickering*, 1903]. Studies intensified in the 1960s when high-resolution Lunar Orbiter images became available [e.g., *Cameron,* 1964; *Kuiper et al.*, 1966]. Theories on the genesis of the lunar sinuous rilles have been wide-ranging, including (1) water erosion channels [e.g., *Pickering*, 1903; *Urey*, 1967], (2) products of volcanic ash flows [*Cameron*, 1964], (3) tectonic features [*Quaide*, 1965], (4) lava drainage channels [*Kuiper et al.*, 1966], (5) collapsed lava tubes [*Greeley*, 1971a, b], and (6) lava channels incised into the substrate by thermal erosion [*Hulme*, 1973; *Carr*, 1974]. When analysis of samples returned during the Apollo program demonstrated the anhydrous nature of the lunar surface and the volcanic origins of the lunar maria, it became clear that the genesis of the rilles was the result of volcanic processes. Thus hypotheses 4, 5, and 6 above remain the primary candidate processes for the formation of the lunar sinuous rilles.

The potential of thermal erosion to produce deep lava channels on the Moon was first assessed by *Hulme* [1973, 1982] and *Carr* [1974]. Their mathematical modeling showed that low-viscosity lunar lavas could have been capable of thermal erosion under both turbulent [*Hulme*, 1973, 1982] and laminar [*Carr*, 1974; *Hulme*, 1982] flow regimes, assuming sustained flow at high rates. They emphasized the qualitative rather than quantitative results of their models, however, owing to uncertainties in some thermal-physical properties such as heat transfer coefficient, viscosity, etc. Later, work by *Head and Wilson* [1981], *Coombs et al.* [1988], and *Tomkinson and Wilson* [1991] further investigated the formation of sinuous rilles, the first applying Hulme's model to additional rilles for which topographic data became available, the second applying Hulme's model to the specific case of Rima Mozart on the lunar nearside, and the third adapting Hulme's technique to investigate the effect of down-rille lava density variations. Their results basically supported the initial conclusions of Hulme and Carr that thermal erosion could have had a significant role in the formation of the lunar sinuous rilles.

The purpose of this work is to reevaluate the role of thermal erosion by lunar lavas as a mechanism for the formation of the sinuous rilles. We have adapted the thermal-fluid dynamic-geochemical model of *Williams et al.* [1998], which was used to examine the erosional potential of Archean komatiite lavas at Kambalda, Western Australia, to model the emplacement of turbulent, low-viscosity lavas on the Moon. Key refinements of this model over that of previous work include (1) the linkage of the physics of lava emplacement, thermal erosion, and assimilation of substrate to the geochemical evolution of the flow; (2) the inclusion of algorithms and equations to calculate the temperature- and composition-dependent thermal, rheological, and fluid dynamic properties of the lava and substrate; (3) the use of a convective heat transfer coefficient more appropriate for fluid lava with flow properties that vary

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during emplacement; and (4) consideration of the effect of vesicles on the physical properties of the flow. Our goal is to assess the importance of thermal erosion as a possible mechanism of sinuous rille formation as part of the overall thermal, fluid dynamic, and geochemical evolution of flowing lunar lavas.

2. Background

2.1. Lunar Sinuous Rilles

The resemblance of the lunar sinuous rilles to terrestrial rivers was noted by some of the earliest workers [*Neison*, 1876; *Elger*, 1895; *Pickering*, 1903; *Firsoff*, 1960]. A good summary of the morphological characteristics of ~130 nearside rilles was given by *Schubert et al.* [1970], who concluded that the morphology of the features required their origin from surface water erosion. As the anhydrous nature of the lunar surface and the volcanic origins of the lunar maria became clear during the era of the Apollo landings (1969-1972), studies of the lunar sinuous rilles focused on their genesis as flow conduits of low-viscosity lavas [e.g., *Kuiper et al.*, 1966; *Oberbeck et al.*, 1969; *Greeley*, 1971a, b; *Murray*, 1971; *Cruikshank and Wood*, 1972; *Howard et al.*, 1972; *Gornitz*, 1973; *Hulme*, 1973; *Carr*, 1974].

The nearside lunar sinuous rilles have lengths ranging from 4 to 340 km (median of 34 km), widths ranging from 0 to 4 km (median of 400 m), width to length ratios ranging from 0.0005 to 0.1 (median of 0.01), depths ranging from 50 to 600 m (although *Schubert et al.* [1970] noted that shadow lengths were indicative of true depths in only 20% of the rilles they studied), depth to width ratios ranging from 0.03 to 0.2, and meander wavelengths ranging from <1 to 7 km (median of 1.7 km) [*Schubert et al.*, 1970]. Sinuous rilles are found mostly on mare surfaces, although there are rare examples of rilles cutting highland materials. The presence of goosenecks, well-developed meanders, rare distributary patterns (Figure 1), narrow meandering channels in wider channels (e.g., Schroeter's Valley), and indications of aggradational materials at the mouths of some rilles [*Schubert et al.*, 1970] all suggest that the sinuous rilles must have formed by emplacement of very low viscosity lavas. However, the role of lava erosion versus tube construction and subsequent roof collapse is still under debate. For example, some rilles have morphological characteristics similar to partially collapsed lava tubes [*Oberbeck et al.*, 1969]. Furthermore, a detailed comparison of Hadley Rille to terrestrial lava tubes [*Greeley*, 1971a; *Spudis et al.*, 1988] suggested that Hadley formed primarily as a constructional feature along preexisting structural depressions and that thermal or mechanical erosion had only a minor role (if any) in the rille's formation. Thus, because of this continuing debate, a reassessment of the potential of thermal erosion by lunar lavas is justified.

2.2. Thermal-Physical Properties of Lunar Lavas

The thermal and rheological properties of lunar mare lavas were first reported by *Murase and McBirney* [1970a, b, 1973] on the basis of experimental work performed on a synthetic lava derived from the composition of Apollo 11 basalts. *Murase and McBirney* [1970a] reported that their lunar lava analog had a high liquidus temperature (1300°-1380°C) and a low dynamic viscosity (0.45-1 Pa s^{-1}). We have used the algorithm of *Shaw* [1972] to plot the liquid viscosities versus temperature (Figure 2) of several lava compositions (Table 1). The viscosity of lunar lavas (2.3 Pa s^{-1} at 1200°C) is more than an order of magnitude lower than that of modern, terrestrial tholeiitic basalt lavas (53.8 Pa s^{-1} at 1200°C), more similar to ancient Precambrian komatiitic lavas (e.g., 4.4 Pa s^{-1} at 1200°C for an 18% MgO komatiitic basalt). This low viscosity is attributed to the lower silica and alumina contents and higher iron and titanium contents of lunar basalts relative to terrestrial basalts.

Murase and McBirney [1970b, 1973] also studied the thermal conductivity of their synthetic lunar sample, and they reported that its thermal conductivity in the melting range (1100°-1380°C) was about half that of a molten terrestrial tholeiitic basalt (~0.63 J m^{-1} s^{-1} °C^{-1} for the lunar lava versus ~1.3 J m^{-1} s^{-1} °C^{-1} for the terrestrial basalt lava). We have plotted the *Murase and McBirney* [1973] temperature-dependent thermal conductivity data with those of several other high-temperature silicate materials [*Birch and Clark*, 1940; *Huppert and Sparks*, 1985; *Snyder et al.*, 1994; *Büttner et al.*, 1998] (Figure 3), and applied several curve fits to the best data to derive equations for thermal conductivity as a function of temperature for use in our lava emplacement model. We favor the use of the exponential equation in Figure 3, because of the controversy about the reliability of the Snyder et al. data [*Shore*, 1995; *Snyder et al.*, 1995]. These results suggest that the thermal conductivity of high-temperature, molten silicates may be lower (~0.4-0.6 J m^{-1} s^{-1} °C^{-1}), perhaps up to an order of magnitude lower, than terrestrial basalts (~1-4 J m^{-1} s^{-1} °C^{-1}). This is important because, as *Murase and McBirney* [1970b] concluded, a crust once formed on very fluid lunar lavas should act as an effective insulator of heat that might enable or enhance long-distance flow.

Hulme [1973, 1982] suggested that if lunar lavas were low-viscosity fluids relative to terrestrial lavas, as supported by the work of *Murase and McBirney* [1970b], then they should have been capable of turbulent emplacement. We have tested this hypothesis by taking typical viscosities and densities inferred for a variety of lava compositions (Table 2) and iteratively solving the following three equations for a range of flow thicknesses:

$$u = \sqrt{\frac{4gh\sin(\psi)}{\lambda}} \tag{1}$$

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

$$Re = \frac{\rho h u}{\mu}, \tag{3}$$

in which u is flow velocity, λ is friction coefficient, and Re is Reynolds number (a complete list of symbols is given at the end of this paper). Equation (1) is for a gravity current moving down a slope and is modified from *Britter and Linden* [1980] and *Jarvis* [1995]. We assume a shallow slope (~0.1°). Equation (2) is a friction coefficient for pipe flow from *Kakaç et al.* [1987]. Equation (3) is a general Reynolds number expression, and assumes that the flow thickness h is the characteristic length. As Figure 4 shows, it is clear that low-viscosity lava compositions such as komatiites, carbonatites, and lunar basalts have the potential for turbulence in very thin flows (0.3-3 m), whereas terrestrial basalts require thicker flows (≥20 m) before turbulence is attained. *Hulme* [1973] suggested turbulent flow was required for large-scale thermal erosion to produce the lunar sinuous rilles; by this logic, the

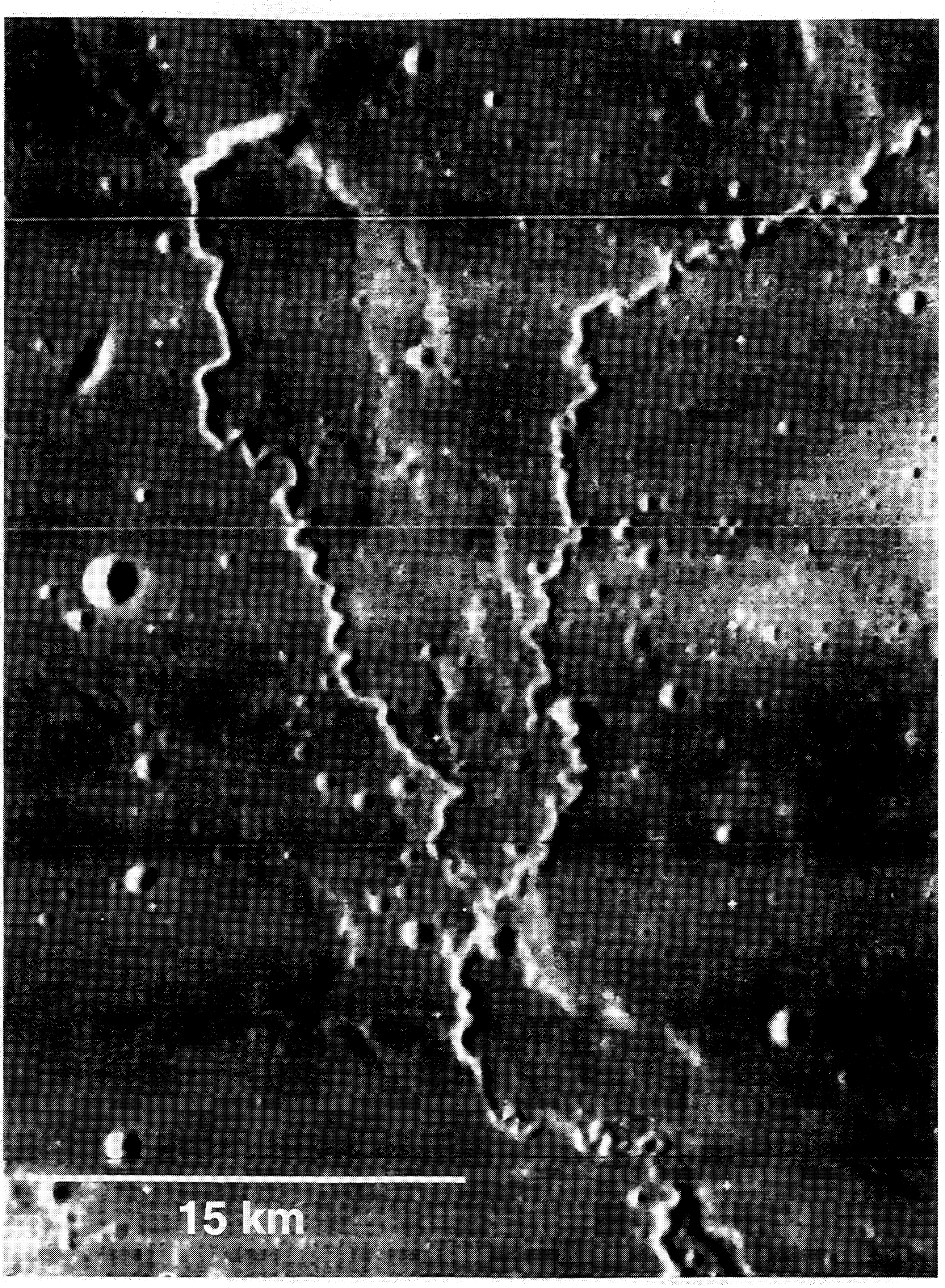

Figure 1. Lunar sinuous rilles northwest of crater Gassendi on the lunar nearside. These rilles have a well-developed meander pattern and show the confluence of two rilles. From *Schubert et al.* [1970]. Lunar Orbiter IV, frame H-137.

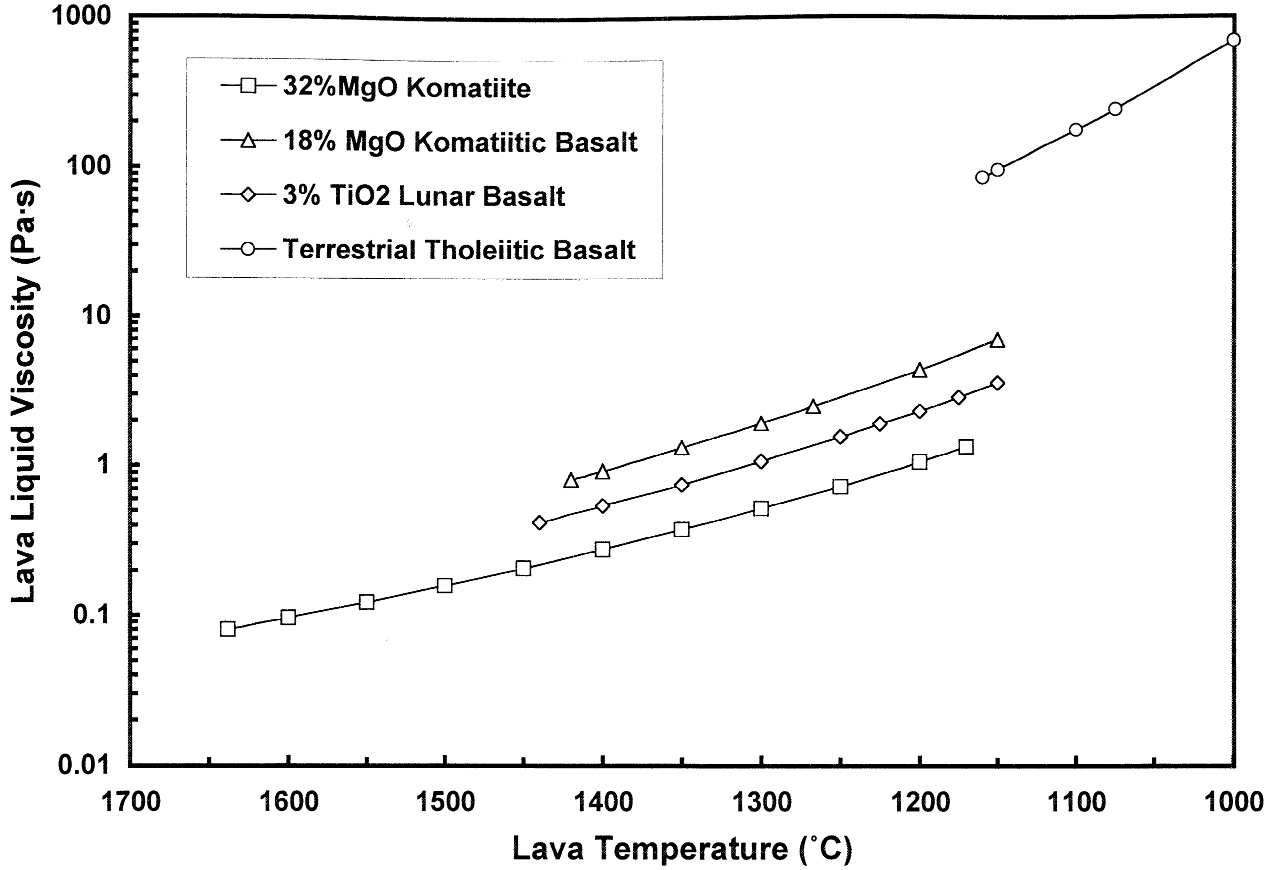

Figure 2. Graph of lava liquid viscosity versus temperature for several lava compositions (Table 1). We used the algorithm of *Shaw* [1972] to calculate lava viscosity as a function of temperature and composition to compare the viscosities of lunar and terrestrial basalts with komatiites. As originally suggested by *Murase and McBirney* [1970a], lunar mare lavas have much lower viscosities than modern tholeiitic basalts, but not as low as ancient komatiites. The closest terrestrial analogs to lunar lavas are low (12-18 wt%) MgO komatiitic basalts, which erupted with komatiites in the Archean but are found to a greater extent in Proterozoic rocks (2.5-0.6 Ga).

onset of turbulent flow can occur in lunar basalt flows of even modest thickness and may be more typical here than in most terrestrial basalts. Hence thermal erosion may be more common on the Moon than is suggested by terrestrial analogs.

2.3. Thermal Erosion

Thermal erosion is the breakup and removal of substrate by hot flowing lava. Strictly thermal erosion involves ablation (i.e., melting) of substrate, although some workers [e.g., *Williams et al.*, 1998] have investigated thermomechanical erosion, which includes some form of physical degradation of loosely consolidated substrate materials by the moving lava. In either case, this erosion may be followed by partial or complete assimilation of melted substrate by the liquid lava. Thermal erosion has been inferred to occur in carbonatite lavas [*Dawson et al.*, 1990], in industrial sulfur flows [*Greeley et al.*, 1990], in terrestrial lava tubes [e.g., *Cruikshank and Wood*, 1972; *Greeley and Hyde*, 1972; *Peterson and Swanson*, 1974; *Peterson et al.*, 1994; *Kauahikaua et al.*, 1998; *Greeley et al.*, 1998], and in Precambrian komatiite flows [e.g., *Huppert et al.*, 1984; *Huppert and Sparks*, 1985; *Groves et al.*, 1986; *Barnes et al.*, 1988; *Lesher*, 1989; *Perring et al.*, 1995; *Williams et al.*, 1998]. Thermal erosion has also been suggested as a possible mode of origin for some Martian lava channels [*Carr*, 1974; *Cutts et al.*, 1978; *Baird*, 1984; *Wilson and Mouginis-Mark*, 1984] and some Venusian canali [*Head et al.*, 1991; *Baker et al.*, 1992; *Komatsu et al.*, 1993; *Komatsu and Baker*, 1994; *Bussey et al.*, 1995]. Most terrestrial evidence of thermal erosion is limited to either field relationships indicative of downcutting by tube-fed basaltic lavas or geophysical measurements indicative of downcutting by active tube-fed lavas. These measurements suggest erosion rates of basalt substrate by laminar, tube-fed basalt lava of up to 10 cm d^{-1} [*Kauahikaua et al.*, 1998], which for sustained tube-fed eruptions of weeks to months suggests maximum erosion depths of >>2-3 m on moderate slopes.

3. Model and Assumptions

The emplacement and thermal erosion model for low-viscosity, turbulently flowing, confined submarine komatiite lavas was summarized by *Williams et al.* [1998] and has been adapted here to study lunar lava flows erupted in a vacuum. Here we describe the primary algorithms of the model. A list of symbols is included at the end of the paper.

We begin by choosing a starting lava major oxide composition. We have chosen to use lunar sample 12002 (Table 1), an Apollo 12 low-titanium picritic basalt [*Walker et* *al.*, 1976], as our model lunar lava composition for several reasons. First, this composition has a high liquidus temperature (~1440°C) and low dynamic viscosity (0.75 Pa s^{-1}), which suggests it should have a high thermal erosive potential. Second, olivine is the only crystallizing silicate phase over most of the turbulent emplacement regime of this lava (clinopyroxene arrives on the liquidus at ~1200°C, after the lava has become laminar), which can be simulated with our model. We have verified that olivine is the sole crystallizing silicate phase by running a fractional crystallization model on this composition using MELTS [*Ghiorso and Sack*, 1995]. The Fe-Ti oxide phase ilmenite also crystallizes during flow, but it is a minor phase (~7 vol% [*Basaltic Volcanism Study Project*, 1981]), and its presence would not strongly affect the lava rheology. Third, this lunar lava has a composition and physical properties somewhat similar to terrestrial komatiitic basalts (Table 1), and thus model results using this composition can be compared to model results for the Cape Smith komatiitic basalts [*Williams et al.,* 1999], which are also inferred to have produced thermal erosion channels in the Precambrian [e.g., *Lesher and Thibert*, 2000].

Next, we use a series of algorithms to calculate initial values of important temperature- and composition-dependent thermal-physical properties of the lava. The liquidus temperature is calculated using MELTS [*Ghiorso and Sack*, 1995], and the solidus temperature is estimated for all compositions. The MELTS calculations used a redox state with the oxygen fugacity (fO_2) set at the Iron-Wüstite buffer. Liquid lava density is calculated using the method of *Bottinga and Weill* [1970] with the partial molar volume data of *Mo et al.* [1982]. Liquid lava dynamic viscosity is calculated using the method of *Shaw* [1972]. Lava specific heat is calculated from the heat capacity data of *Lange and Navrotsky* [1992]. Lava heat of fusion is approximated using curve fits to the data of *Navrotsky* [1995]. Lava thermal conductivity is calculated using the exponential equation from Figure 3.

We make several assumptions about other relevant parameters. The substrate composition is assumed to be similar to the overlying lava, which is consistent with Lunar Orbiter photographs of most rilles. The initial lava flow thickness is assumed to be 10 m (Table 3), which is consistent with previous estimates of the thicknesses of individual lunar flows [*Schaber*, 1973; *Brett*, 1975; *Schaber et al.*, 1976; *Gifford and El-Baz*, 1981]. We assume no inflation of the lava during emplacement. The lavas are assumed to erupt in a vacuum and to flow over a flat, homogeneous, massive basaltic substrate with a shallow slope of ~0.1°.

From the initial values of the lava thermal-physical properties, a series of auxiliary equations is used to calculate additional lava properties at the vent (where lava temperature $T=T_0$, the eruption temperature), and subsequently at increments of distance downstream. First, lava crystallinity X is given by the ratio of the degree of undercooling divided by the range of crystallization:

$$X = \frac{T_{liq} - T}{T_{liq} - T_{sol}}; \tag{4}$$

Table 1. Inferred Liquid Compositions and Physical Properties for Several Komatiitic and Basaltic Lavas

Component	High-MgO Komatiite	Low-MgO Komatiitic Basalt	Low-TiO_2 Lunar Mare Basalt	High-TiO_2 Lunar Mare Basalt	Synthetic Lunar Mare Basalt	Terrestrial Tholeiitic Basalt
SiO_2	45.0	46.9	43.6	38.6	43.0	50.9
TiO_2	0.3	0.6	2.6	8.8	11.0	1.7
Al_2O_3	5.6	9.8	7.9	6.3	7.7	14.6
Fe_2O_3	1.4	-	-	-	-	-
FeO	9.2	14.4	21.7	22.0	21.0	14.6
MnO	0.2	0.3	0.3	-	0.3	-
MgO	32.0	18.9	14.9	14.4	6.5	4.8
CaO	5.3	8.6	8.3	7.7	9.0	8.7
Na_2O	0.6	0.3	0.2	0.4	0.4	3.1
K_2O	0.03	0.05	0.05	0.09	0.2	0.8
T_{liq}, °C	1638	1419	1440	1338	1166	1160
T_{sol}, °C	1170	1150	1150	1130	1050	1080
ρ at T_{liq}, kg m^{-3}	2770	2800	2900	2990	2980	2750
c, J kg^{-1} $°C^{-1}$	1790	1640	1570	1565	1460	1480
μ at T_{liq}, Pa s^{-1}	0.08	0.74	0.40	0.24	4.2	86
L at T_{liq}, J kg^{-1}	6.97E+05	5.96E+05	6.06E+05	5.87E+05	5.39E+05	5.37E+05
k at T_{liq}, J m^{-1} s^{-1} $°C^{-1}$	0.4	0.5	0.5	0.5	0.6	0.7
h, m	10	10	10	10	10	10
u at T_{liq}, m s^{-1}	4.8	3.9	4.2*	4.3*	3.0*	1.7
Re at T_{liq}	1.74E+06	1.46E+05	3.09E+05*	3.79E+05*	1.52E+04*	5.50E+02
Pr at T_{liq}	3.8E+02	2.6E+03	1.3E+03*	1.0E+03*	1.4E+04*	1.9E+05
h_T at T_{liq}, J m^{-1} s^{-1} $°C^{-1}$	396	172	217*	246*	73*	N/A
u_{mb} at T_{liq}, m d^{-1}	1.7	0.52	0.24*	0.39*	0.004*	N/A
Composition location	Kambalda, Western Australia	Katinniq, Cape Smith Belt, Canada	Apollo 12 Sample 12002	Apollo 17 Sample 74220	SLS, based on Apollo 11 Basalts	CRB, Washington
Composition reference	[1]	[2]	[3]	[3]	[4]	[4]

Symbols: 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; N/A, not applicable for nonturbulent flows. Notes: For all compositions, liquidus temperature was calculated using MELTS [*Ghiorso and Sack*, 1995]. The solidus temperature for komatiite is from *Arndt* [1976] and is estimated for all other compositions. Liquid density was calculated using the method of *Bottinga and Weill* [1970] with the partial molar volume data of *Mo et al.* [1982], liquid viscosity was calculated using the method of *Shaw* [1972], specific heat was calculated from the heat capacity data of *Lange and Navrotsky* [1992], and heat of fusion for komatiite liquids is approximated using curve fits to the data of *Navrotsky* [1995]. Thermal conductivity was calculated using the exponential equation from Figure 3, which is a curve fit to the conductivity data of *Murase and McBirney* [1973] and *Snyder et al.* [1994]. Flow velocity calculated using equation (1), Reynolds number by equation (7), Prandtl number by equation (8), heat transfer coefficient by equation (9), and erosion rate by equation (10). References: 1, Adapted from *Lesher and Arndt* [1995]; 2, *Barnes et al.* [1982]; 3, *Walker et al.* [1976]; 4, *Murase and McBirney* [1973].
*If erupted on Earth.

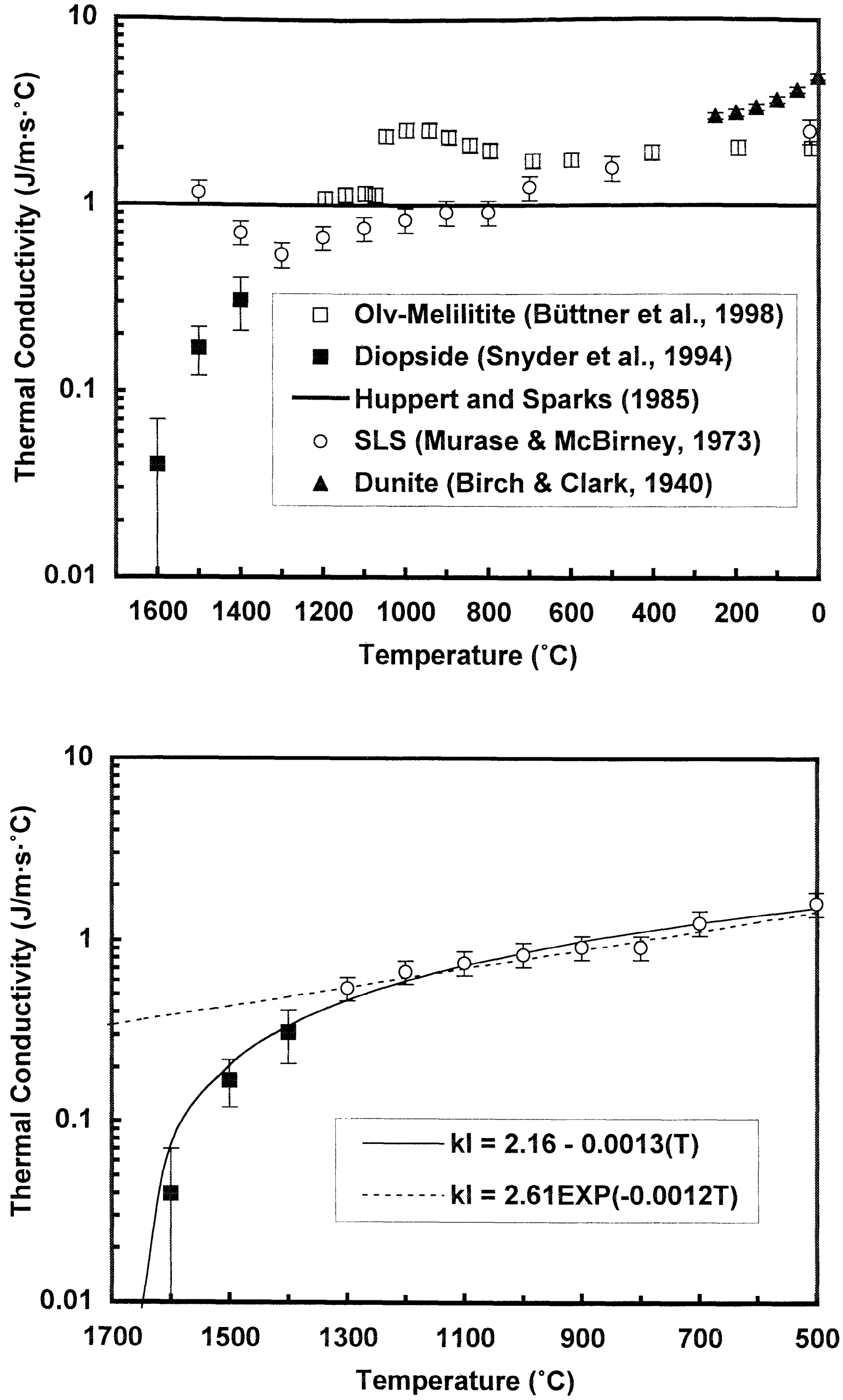

Figure 3. (top) Graph of thermal conductivity versus temperature for mafic and ultramafic rocks and liquids from several data sets [*Birch and Clark*, 1940; *Murase and McBirney*, 1973; *Snyder et al.*, 1994]. (bottom) Two curve fits to the best data which provide equations for calculating thermal conductivity as a function of temperature. Because of the controversy about the reliability of the Snyder et al. data [*Shore*, 1995; *Snyder et al.*, 1995], we use the exponential equation (dashed line) in our model.

Table 2. Typical Viscosities and Densities Inferred for Various Lava Compositions

Lava	Viscosity, Pa s^{-1}	Density, kg m^{-3}	Reference
High-MgO komatiite	0.1	2800	*Huppert and Sparks* [1985]
Carbonatite	1	2200	*Norton and Pinkerton* [1992]; *Dawson et al.* [1990]
Lunar basalt	2	2900	*Murase and McBirney* [1970b, 1973]
Terrestrial basalt	100	2800	*Turcotte and Schubert* [1982]
Terrestrial andesite	1000	2500	*Williams and McBirney* [1979]
Terrestrial rhyolite	10,000,000	2300	*McBirney* [1993]

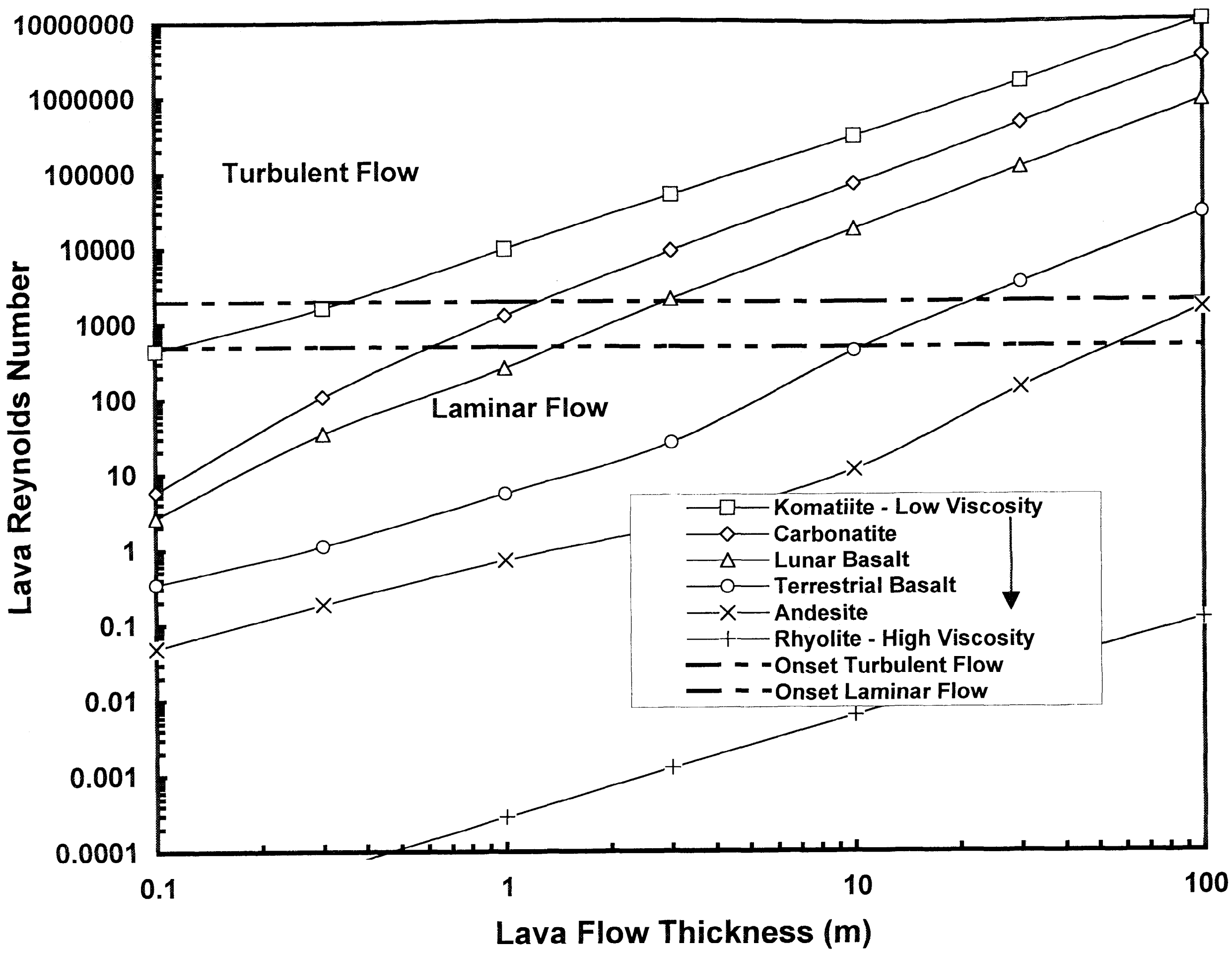

Figure 4. Graph of Reynolds number versus flow thickness for a variety of lava compositions. Komatiites, carbonatites, and lunar basalts have the potential to erupt turbulently even as thin flows (≤1-2 m thick), owing to their low viscosities. In contrast, terrestrial tholetiitic basalts are limited to laminar flow unless erupted as thick flows (≥10 m thick).

in which T_{liq} is liquidus temperature, T_{sol} is solidus temperature, and T is lava temperature at this model increment. This is a simplistic method of calculating crystallinity, in that it assumes constant liquidus and solidus temperatures rather than a range of temperatures. Equation (4) also assumes a linear growth in crystal fraction with cooling; although this is not valid for lavas undergoing crystallization of multiple silicate phases (e.g., terrestrial basalts), we have verified using MELTS that this equation gives an adequate approximation for lavas crystallizing a single silicate phase (olivine) for the cooling range of interest.

Bulk viscosity μ_b when $X < 0.3$ is given by

$$\mu_b = \mu_l \left(1 - \frac{X}{0.6}\right)^{-2.5}, \tag{5}$$

in which μ_l is liquid viscosity, and when $X \geq 0.3$ is given by

$$\mu_b = \mu_l \exp\left\{\left[2.5 + \left(\frac{X}{0.6 - X}\right)^{0.48}\right]\frac{X}{0.6}\right\} \tag{6}$$

Equation (5) is the well-known Roscoe-Einstein equation, and

Table 3. Input Values for Modeling Low-Viscosity Lava Emplacement on the Moon and Earth

Parameter	Moon	Earth
Lava composition (Table 2)	Apollo 12 sample 12002	Cape Smith komatiitic basalt
Eruption temperature, °C	1440	1419
Liquidus temperature, °C	1440	1419
Solidus temperature, °C	1150	1150
Flow thickness, m	10	10
Emplacement environment	vacuum	submarine
Ambient temperature, °C	0	0
Heat loss from upper surface	radiation	convection
Crystallizing phases	100% olivine	100% olivine
Substrate composition	lunar basalt, tholeiitic basalt	komatiitic basalt, tholeiitic basalt
Substrate slope	0.1°	0.1°
Gravity, m s^{-2}	1.622	9.812

(6) is from *Pinkerton and Stevenson* [1992] and is recommended for calculating bulk viscosity at high crystal fractions. Equation (6) considers only the viscosity effect of liquid and entrained crystals, not the effect of the overlying crust. While we acknowledge that additional factors such as the strength of the crust may influence the bulk motion of the flow, for the purpose of this model we assume that crystal growth (and presence of vesicles, which we discuss later) in the lava interior is the only important parameter in calculating bulk viscosity.

We next calculate flow velocity u, friction coefficient λ, and Reynolds number Re iteratively using (1) - (3), except that (3) is substituted with

$$Re = \frac{\rho_l h u}{\mu_b}, \tag{7}$$

in which μ_b is bulk viscosity from either (5) or (6), as appropriate. All model results are reported down to Re = 2000, the theoretical limit on turbulent flow in conduits (channels or tubes). The lava Prandtl number Pr is given by

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

in which c_l is specific heat and k_l is thermal conductivity.

Hulme [1973] and *Huppert and Sparks* [1985] showed that low-viscosity, turbulently flowing lavas should loose heat to the surroundings primarily by convection rather than conduction. Both used a convective heat transfer coefficient for turbulent pipe flows for fluids with constant physical properties. However, lava flows not only undergo phase changes but also have temperature-dependent physical properties. Thus a more appropriate convective heat transfer coefficient h_T is desirable, which includes the effect for fluids with temperature-dependent physical properties. Such an expression for turbulent pipe flows [see *Kakaç et al.*, 1987] is given by

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

in which k_{eff} is the effective thermal conductivity in the thermal boundary layers at the top and base of the flow and μ_g is the viscosity of the melted substrate. The ratio of the lava bulk viscosity to the viscosity of the melted substrate in (9) typically has the effect of reducing heat transfer compared to that found in fluids with constant fluid physical properties.

The lava erosion rate u_m, as modified from *Huppert and Sparks* [1985], is given by

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

in which T_{emg} is the effective melting temperature of the substrate (which for a given value of μ_g maximizes thermal erosion) and E_{mg} is the energy required to melt the substrate, which is given by

$$E_{\mathrm{mg}} = \rho_{\mathrm{g}} \left[c_{\mathrm{g}} \left(T_{\mathrm{mg}} - T_{\mathrm{a}} \right) + f_{\mathrm{L}} L_{\mathrm{g}} \right], \tag{10b}$$

in which ρ_g is substrate density, c_g is substrate specific heat, T_{mg} is substrate melting temperature, T_a is ambient temperature of the surface, and L_g is heat of fusion required to melt the substrate. The erosion rate is used to calculate the degree of contamination of the lava by assimilated substrate S [*Williams et al.*, 1998], given by

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

in which Q_0 is the initial flow rate and $Q(x)$ is the flow rate at any given distance from the source.

The compositional change in the liquid lava due to the assimilation of thermally eroded substrate S at each model distance increment is calculated using the following mass balance expression:

$$M_{\mathrm{new}} = M_{\mathrm{old}} (1 - \Delta S) + M_{\mathrm{asm}} (\Delta S), \tag{12}$$

in which M_{new} is the major oxide composition of the lava at the current distance from the source, M_{old} is the major oxide composition of the lava at the previous distance increment, and M_{asm} is the major oxide composition of the substrate. Similarly, the compositional change in the liquid lava due to the crystallization of olivine X can also be determined using the following mass balance expression:

$$M_{\mathrm{new}} = M_{\mathrm{old}} (1 - \Delta X) - M_{\mathrm{olv}} (\Delta X) \tag{13}$$

in which M_{old} in (13) is M_{new} from (12) and M_{olv} is the olivine major oxide composition at the current distance from the source. Equation (13) is used in conjunction with partition coefficient and stoichiometric algorithms to calculate M_{olv} at each model increment. Olivine-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]. Equation (13) results in a new liquid lava composition (M_{new}) that is used to recalculate the temperature- and composition-dependent thermal, rheological, and fluid dynamic properties of the lava at each model increment.

The key parameter that advances our model is the lava temperature, which decreases with each increment of distance from the source. As shown in Table 1, lunar lavas are inferred to have been of much higher temperature and lower viscosity than modern terrestrial basaltic lavas, more like terrestrial komatiitic lavas. We thus apply our model for komatiite lava emplacement, assuming that (1) lava erupts as a turbulent flow with a thermally mixed interior (i.e., no thermal or compositional heterogeneities or velocity variations across the width or the depth of the flows), (2) convective rather than conductive heat tranfer occurs between the flow interior and the top of the flow (maintained at the lava solidus temperature with a coherent crust) and base of the flow (maintained at the substrate melting temperature), (3) thermal erosion occurs at the base of the flow when the lava temperature is greater than the melting temperature of the substrate, and (4) latent heat is released as the flow crystallizes [see also *Huppert and Sparks*, 1985]. We further assume (for simplicity) one-dimensional flows (flow in the x direction) with thermal erosion in the z direction and that these lavas were erupted as single flow units at a conserved flow rate, in which flow thickness increases as flow velocity decreases. This model of lava cooling with distance, including convective heat transfer, thermal erosion, and release of latent heat, is given by the following first-order ordinary differential equation modified from *Huppert and Sparks* [1985]:

$$\rho_b c_l h u \frac{dT}{dx} = -h_T\left(T - T_{mg}\right) - h_T\left(T - T_{sol}\right) - \frac{\rho_b c_l h_T \left(T - T_{mg}\right)^2}{E_{mg}} + \rho_b c_l h u \frac{dT}{dx}\frac{L_l x^{'}(T)}{c_l} \quad (14)$$

Because the physical properties of the lava are changing with distance, (14) must be solved at each increment of distance from the eruption source using a fourth-order Runge-Kutta numerical method. Once a new temperature (from equation (14)) and a new lava composition (from equation (13)) are calculated, all of the important thermal, rheological, and fluid dynamic parameters are calculated at that distance. Thus the physical and geochemical evolution of the lava flow with distance is simulated.

Another important consideration for lava emplacement on the Moon and Earth is the formation of a surface crust. On Earth, field evidence suggests subaqueous emplacement of komatiitic flows at all localities [e.g., *Lesher*, 1989], and thus a glassy crust should have formed quite quickly upon quenching in cold seawater. This crust would have quickly covered the flow, providing a stable "insulating boundary" that enhanced long-distance flow. At present there is some debate about whether komatiites had an initial crust (if komatiites erupted in a superheated state, a crust would not form until lava temperature dropped below the liquidus), about what part of preserved komatiite exposures could correspond to a crust, and about what the nature of a crust should be over a turbulent flow (fragmented or not). *Crisp and Baloga* [1994] provide some ways of quantifying a disrupted crust. For simplicity, *Williams et al.* [1998, 1999] assumed an undisrupted crust whose growth was constrained by convective cooling of overlying seawater. We assume the same undisrupted crust for the lunar flows, except that on the Moon crustal growth is constrained by radiative cooling to a vacuum, in which the rate of crustal growth (dh_c/dt) is given by

$$\frac{dh_c}{dt} = u\frac{dh_c}{dx} = \frac{\left[\varepsilon\sigma\left(T_c^4 - T_a^4\right)\right] - h_T\left(T - T_{sol}\right)}{\rho_b\left[c_l\left(T - T_{sol}\right) + L_l\right]}, \quad (15)$$

in which ε is lava emissivity, σ is Stefan-Boltzmann radiative constant, and T_c is the contact temperature of the lava with the ambient environment. Now at some distance from the vent a balance is achieved between the heat fluxes in (15). 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:

$$\varepsilon\sigma\left(T_{cs}^4 - T_a^4\right) = h_T\left(T - T_{sol}\right). \quad (16a)$$

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

$$T_{cs} = \sqrt[4]{T_a^4 + \frac{h_T\left(T - T_{sol}\right)}{\varepsilon\sigma}}. \quad (16b)$$

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

$$h_{cs} = \frac{k_c\left(T_{sol} - T_{cs}\right)}{\varepsilon\sigma\left(T_{cs}^4 - T_a^4\right)}, \quad (17)$$

in which k_c is thermal conductivity of the crust. Note that these steady state equations are useful only on long time scales, and cannot be used to study short-term variations in the flow. This calculation allows us to estimate roughly the thickness of the crust on the flow at any given distance from the eruption source on the basis of the heat loss from the turbulent lava into the cold lunar environment.

There is considerable evidence to suggest that some lunar lavas had high volatile gas contents (e.g., vesicular basalt samples returned during the Apollo program, presence of pyroclastic deposits). Thus the effect of vesicles on the emplacement of lunar lavas should be assessed. The presence of bubbles in lava should decrease lava density and specific heat and increase viscosity at low strain rates [*Pinkerton and Stevenson*, 1992]. However, turbulent flow suggests high strain rates, for which the effect of bubbles has not been assessed (L. Keszthelyi, personal communication, 2000). The best that can be done is to evaluate bubble effect on viscosity using available information at low strain rates. 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:

$$\rho_{eff} = f_v\left(\rho_{gas}\right) + \left(1 - f_v\right)\rho_b \quad (18)$$

$$c_{eff} = f_v\left(c_{gas}\right) + \left(1 - f_v\right)c_l \quad (19)$$

$$\mu_{eff} = \mu_b\left[\frac{1}{1 - \left(1.3 f_v\right)^{1/3}}\right], \quad (20)$$

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 plus crystals), and *l* refers to liquid. Equation (20) is from *Sibree* [1934], which is valid for foams with values of f_v up to 75% and which *Pinkerton and Stevenson* [1992] suggest is still a useful algorithm to assess the effect of bubbles on lava viscosity. For simplicity, we assume that f_v is constant during emplacement. In reality, degassing occurs during emplacement, in which bubbles tend to migrate to a free surface and are lost [*Peitersen et al.*, 2000]. *Sato* [1978] suggested that CO was the main gas phase on the Moon [*Heiken et al.*, 1991], and it is the logical choice for our model volatile component. Although very high temperature measurements of the thermal and rheological properties of CO do not exist, the effect of this volatile gas on lava emplacement can still be assessed with available data: ρ_{gas} = 0.793 kg m^{-3} and c_{gas} = 1257 J kg^{-1} K^{-1} at 1500 K [*Lide*, 1993].

4. Results

We ran models for the emplacement of an initially 10 m thick lunar lava flow on the Moon and the Earth to assess planetary environmental effects, and we ran the same models for a Cape Smith komatiitic basalt lava on the Moon and the Earth to assess compositional effects. Results confirm that channelized lunar lavas could have erupted as turbulent flows on the Moon (Figure 5, Table 4). For example, our model lunar

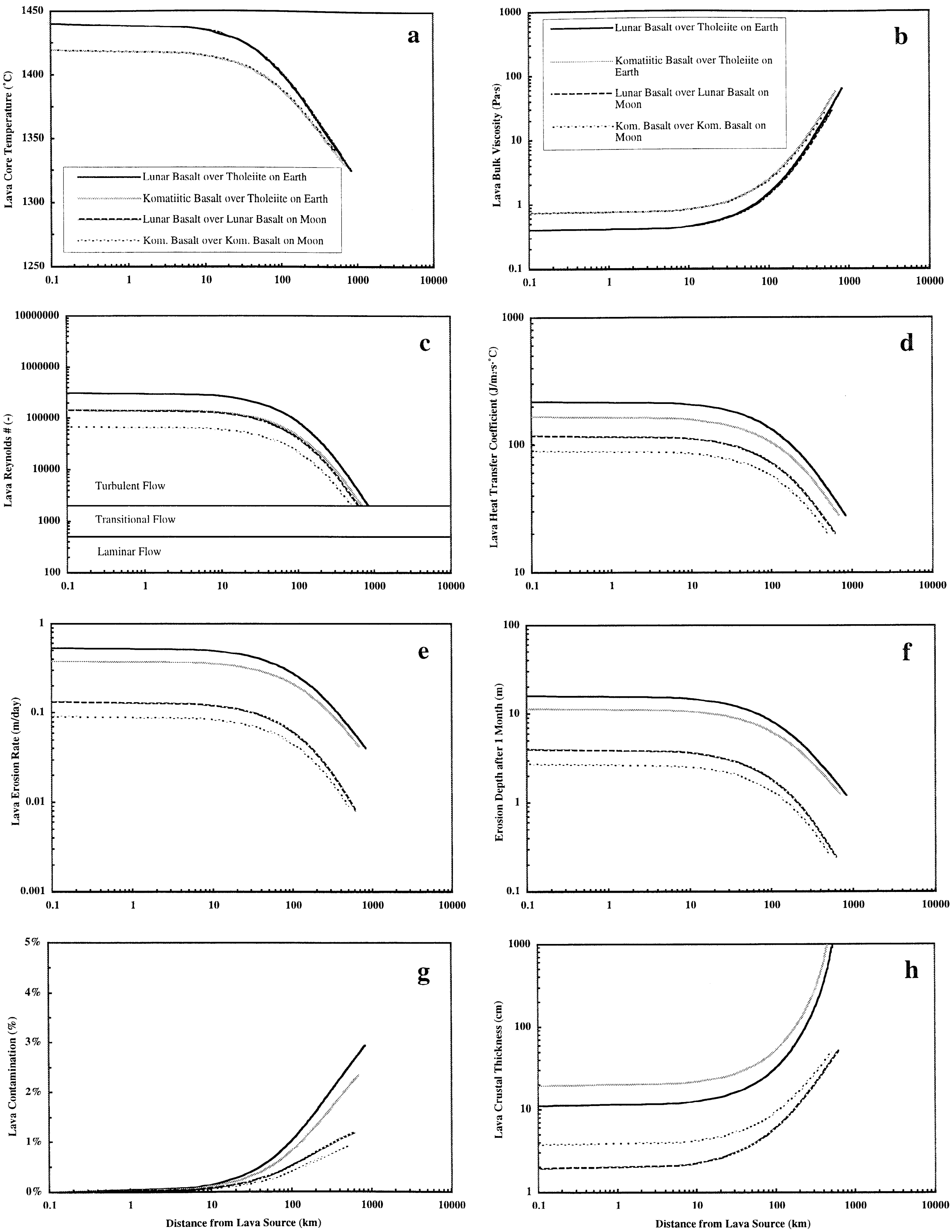

Figure 5. Modeling results for the emplacement of low-viscosity, initially 10 m thick lunar and komatiitic basalt lava flows over basaltic substrate on the Earth and the Moon. Model input parameters are given in Tables 1 and 3; results are given in Table 4.

Table 4. Model Results for the Emplacement of Lunar and Komatiitic Basalt Lava Flows Over Basaltic Substrate on the Earth and the Moon

Property	Lunar Basalt on Earth	Komatiitic Basalt on Earth	Lunar Basalt on the Moon	Komatiitic Basalt on the Moon
Maximum flow distance,* km	835	690	620	500
Maximum flow thickness,* m	15.8	15.0	14.8	14.0
Maximum Reynolds number†	3.1E+05	1.5E+05	1.4E+05	6.8E+04
Maximum Prandtl number†	1.3E+03	2.6E+03	1.3E+03	2.6E+03
Maximum heat transfer coefficient,† J m^{-2} s^{-1} °C^{-1}	220	160	120	89
Substrate composition	tholeiite	tholeiite	lunar basalt	komatiitic basalt
Maximum erosion rate,† cm d^{-1}	53	38	13	9.1
Maximum erosion depth after 1 week,† m	3.7	2.7	0.93	0.64
Maximum contamination,* %	2.9	2.4	1.2	0.9
Maximum crustal thickness,* cm	>1100	>1100	52	50

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

flow could have traveled ~620 km before the turbulent-transitional boundary was reached (*Re* ~ 2000), assuming a sufficiently flat surface and unobstructed flow. This distance is much greater than if a komatiitic basalt lava was erupted under similar conditions on the Moon (~500 km) and is much less than if a lunar lava of this thickness was erupted on Earth (~835 km). There are several competing factors that produced these results. First, because of the lower silica and alumina content and higher iron and titanium content of the lunar basalt lava, it is inferred to have had a higher liquidus temperature (~1440° versus 1420°C), lower dynamic viscosity (0.40 versus 0.74 Pa s^{-1}), and higher density (2900 versus 2800 kg m^{-3}) than the komatiitic basalt lava. Second, the lower lunar gravity relative to Earth (1.6 versus 9.8 m s^{-2}) causes flows of the same thickness to erupt at lower rates and velocities (e.g., ~2.0 m s^{-1} on the Moon versus ~4.2 m s^{-1} on Earth). Although the thermal-rheological factors and gravitational factors compete against each other, the end result is longer (potential) turbulent flow distances and higher Reynolds numbers (greater turbulence) of the lunar lavas versus komatiitic basalt lavas. However, topographic irregularities would likely restrict flow distances to a few hundred kilometers at most, consistent with the lengths of the sinuous rilles (4-340 km [*Schubert et al.*, 1970]).

Most sinuous rilles appear to have formed in the lunar maria, i.e., in basaltic material of the same or nearly the same composition [e.g., *Schubert et al.*, 1970]. Thus we must assess the potential of lunar lavas to thermally erode substrate of a similar composition to the lava. Our results show that the maximum thermal erosion rates for a turbulent lava erupted over a consolidated substrate of the same composition are very low: ~13 cm d^{-1} (at the vent) for the lunar basalt and ~9 cm d^{-1} for the komatiitic basalt. Thus, assuming a "typical" estimated rille depth of 100 m, an initially 10 m thick lunar basalt flow would require a long flow duration of ~770 days (~2.1 years) to produce such a channel near the vent by thermal erosion. In contrast, maximum erosion rates are higher over a substrate that is more silicic with a lower melting temperature, such as a terrestrial tholeiite. On Earth a lunar basalt might have an erosion rate ~53 cm d^{-1} over a tholeiitic substrate, which would require a flow duration of ~188 days (~6.3 months) to produce a 100 m deep channel near the vent. This appears to be a fundamental difference between the Earth and the Moon: on Earth the largest (~100 m deep) interpreted thermal erosion channels occur at localities where the substrate is more felsic (lower melting temperature) than the eroding lava and/or the substrate is loosely consolidated (e.g., Perseverance, Western Australia: komatiite lava over felsic tuff [*Barnes et al.*, 1988; *Williams et al.*, 1999]; Katinniq, New Québec: komatiitic basalt lava over gabbro [*Williams et al.*, 1999]). However, considerable erosion (>20 m) has recently been inferred in Hawaiian lava tubes [*Greeley et al.*, 1998; *Kauahikaua et al.*, 1998], in which the substrate and lava are of the same composition (basalt). Thus it seems likely that thermal erosion is a viable mechanism for genesis of the lunar sinuous rilles, given sufficiently long eruption durations.

If the lunar lava was volatile-rich, and thus contained a significant fraction of vesicles, our modeling (Figure 6) suggests that the reduction in density and specific heat and increase in viscosity that accompany vesiculation would have decreased maximum turbulent flow distances and maximum erosion rates. For example, an initially 10 m thick, moderately vesiculated lunar flow (30% vesicles) would have had a maximum potential turbulent flow distance of 385 km and a maximum thermal erosion rate of 6.2 cm d^{-1}. A heavily-vesiculated flow (50% vesicles) would have had an even lower maximum potential turbulent flow distance (230 km) and maximum thermal erosion rate (3.8 cm d^{-1}). Even higher degrees of vesiculation (>65%) would have resulted in viscosities that inhibited turbulent emplacement.

6. Discussion

Our results support the hypothesis of *Hulme* [1973, 1982] that lunar lavas could have erupted as turbulent flows that had the potential to flow long distances, of the order of hundreds of kilometers. However, our results also show that large-scale thermal erosion is difficult when the substrate is a consolidated basalt of the same composition as the lava. Long-duration eruptions (i.e., months to years) would have been required for thin (~10 m) lunar flows to incise deep (50-100 m) channels in a lunar basaltic substrate by thermal erosion, and thus large volumes of lava would have been transported through the deepening channel. However, the inference of >20 m of basalt erosion by basaltic lava in a Kilauean lava tube during its ongoing eruption (months) [*Kauahikaua et al.*, 1998] may strengthen the likelihood of large-scale thermal erosion on the Moon.

Hulme [1973] suggested that thermal erosion might be enhanced by partial melting of the substrate. Specifically, he suggested that when basalt began to melt at its solidus temperature, only some fraction of the heat of fusion (e.g., f_L = 40%) was required to remove low melting temperature

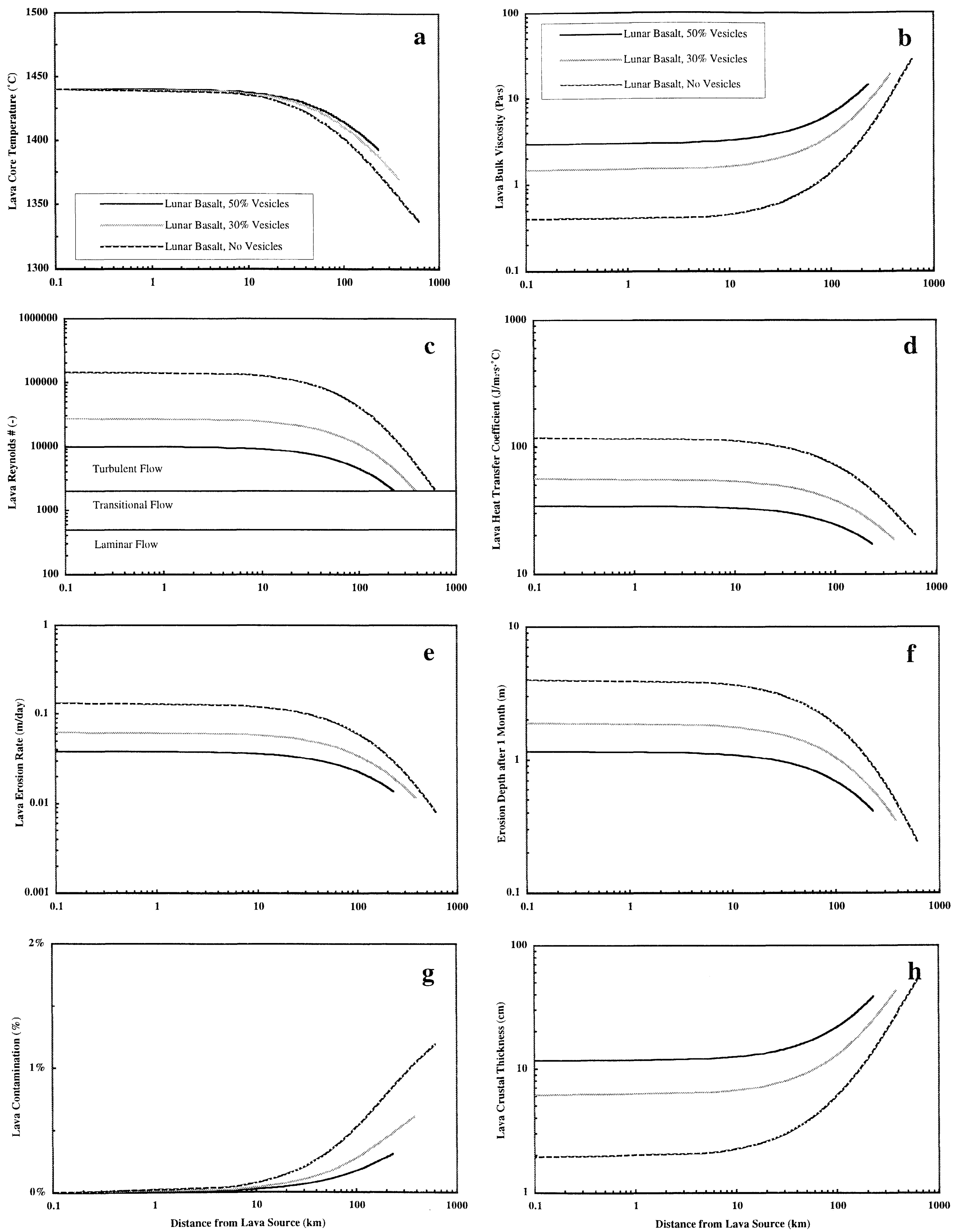

Figure 6. Modeling results for the emplacement of low-viscosity, initially 10 m thick lunar basalt lava flows of various vesicularities over a lunar basalt substrate on the Moon.

materials and that the remaining material could have been rapidly incorporated into the turbulently flowing lava through mechanical erosion. This hypothesis can be tested with our model, and these results are given in Figure 7 and Table 5. Our results show that there is a minor increase in erosion rate when the substrate requires only 40% of the heat of fusion for removal (15 versus 13 cm d^{-1}), which would reduce the time required to produce a channel of a given depth (e.g., ~670 days to produce a 100 m deep channel).

Are there any other factors that might enhance thermal erosion as a viable mechanism for rille formation under these conditions (i.e., lava and substrate of similar composition, consolidated substrate, low slopes)? Higher flow rates (either thicker or faster moving flows) would result in larger erosion rates. For example, an initially 50 m thick lunar basalt flow (Table 4) would have a maximum erosion rate of 22 cm d^{-1}, which would require an eruption duration of 455 days (1.2 years) to produce a 100 m channel. Although studies of the relatively young, well-preserved Imbrium flows show that they are thin (≤10-15 m [*Brett*, 1975; *Schaber et al.*, 1976; *Gifford and El-Baz*, 1981]), they may not be typical of all lunar mare flows.

Another possible way of producing higher thermal erosion rates is by the emplacement of superheated lavas. For example, by superheating the lavas 200°C above their liquidi, it is possible to generate much higher erosion rates over substrates of the same composition (Figure 7; Table 5). In this case, our model predicts that an initially 10 m flow would have an erosion rate of 41 cm d^{-1} at the vent, which would require a flow duration of ~243 days (~8.1 months) to produce a 100 m deep channel. Our results also predict that a superheated lunar lava flow would travel ~65 km in an open channel before radiative cooling would induce crust formation. If superheating occurred during the eruption of lunar lavas, it might explain not only the formation of the sinuous rilles (by thermal erosion) but also why the morphology of most rilles resembles open channels but not closed lava tubes.

How likely was the eruption of superheated lavas on the Moon? This is difficult to assess. On Earth the eruption of superheated lavas has never been observed, and the geological evidence for superheating is uncertain. Theoretically, a magma from a deep planetary source will become superheated if it ascends sufficiently fast that heat loss to the surroundings is minimized. This process requires that the magma ascend rapidly through the lithosphere [*Arndt*, 1994], following a path close to the liquid adiabat that is steeper than the liquidus [e.g., *Kerr et al.*, 1996]. Although some workers suggest that superheating was associated with magma oceans and the hot planetary interiors associated with planetary accretion [e.g., *Jakes*, 1992], others suggest that superheating can occur in mantle plumes and may have produced some of the Cretaceous-aged Gorgona Island komatiite lavas [*Kerr et al.*, 1996] on Earth. Superheating also may have been responsible for the high degrees of lava contamination found in the ~2.7 Ga Perseverance komatiites in Western Australia [*Williams et al.*, 1999]. Thus, if superheated lavas were erupted on Earth both in the Precambrian and in the Phanerozoic, then, in principle, they could have erupted during the time of sinuous rille formation (2.7-3.6 Ga) on the Moon. Recently, in studies of magma buoyancy related to the eruption of mare basalts, *Wieczorek et al.* [2000] suggested, on the basis of geophysical studies and thermal modeling, that mare basalts could have been superheated to 1900 K (~1630°C) at 3.5 Ga in their source region, which enabled the denser lunar basalts to erupt through the lunar crust using buoyancy alone. In light of the *Wieczorek et al.* [2000] work and our model results, perhaps the plausibility of the emplacement of superheated lavas on the Moon should be investigated further.

Although this model is more rigorous than previous models of thermal erosion on the Moon [*Hulme*, 1973; *Carr*, 1974], there are still uncertainties in modeling lunar lava emplacement that need to be addressed. First, our model is designed to study turbulent flow emplacement. Because we now have evidence that laminarly flowing basaltic lava can substantially thermally erode basaltic substrates during long-duration eruptions [*Kauahikaua et al.*, 1998], a more rigorous treatment of the erosive potential of laminarly flowing lunar lavas is in order. Second, it is unclear whether an insulating crust can be sustained on a turbulent flow. Presumably, any crust that formed would be continually fractured, entrained, and reformed (perhaps like those on channelized 'a'a flows [*Keszthelyi and Self*, 1998]), thus limiting its insulating capability. Third, our model does not consider the alterations in flow properties that occur in curves or bends in the lava stream. Presumably, an increased flow velocity (and the erosion rate) at the outer edge of a bend would promote the growth of meanders, perhaps increasing the sinuosity of the rilles. Fourth, the presence of vesicles in lunar basalts may enhance cooling during emplacement, as they appear to do in terrestrial pahoehoe flows [*Keszthelyi*, 1994], which this model does not address.

We are currently studying these issues using the PHOENICS finite-volume computational fluid dynamic software (http://www.cham.co.uk) to solve numerically the fully coupled three-dimensional suite of mass, momentum, and energy conservation equations. With application of realistic geometries, boundary conditions, and temperature-varying thermophysical properties, we are able to explore the detailed interactions between flow and heat transfer in the dynamic and thermal boundary layers at the base of the lava, which are likely to be critical in the process of thermal erosion [*Fagents and Greeley*, 2000]. We thus hope to elucidate further the role of thermal erosion on both the Earth and the Moon.

7. Summary

We have adapted the model of *Williams et al.* [1998] to reevaluate the lava emplacement and thermal erosion potential of low-viscosity lunar lavas as a mechanism of formation of the lunar sinuous rilles. We have modeled lunar lava emplacement using an Apollo 12 low-TiO_2, high MgO basalt composition, which is inferred to have had a high liquidus temperature (1440°C), a high density (2900 kg m^{-3}), and a low viscosity (0.4 Pa s^{-1}) and which had olivine as its sole rheology-altering crystallization phase. We have also modeled the emplacement of a rheologically similar Cape Smith komatiitic basalt lava to assess compositional effects, and we have modeled emplacement on the Moon and the Earth to assess planetary environmental effects. Our results support the hypothesis of *Hulme* [1973, 1982] that low-viscosity lunar basalts could have erupted as turbulent flows on the Moon, which could have been emplaced over hundreds of kilometers given a sufficiently flat and unobstructed substrate. Although the lower lunar gravity results in a lower flow rate (all else equal) that tends to inhibit long-distance flow, the lower SiO_2 and Al_2O_3 contents and higher FeO and TiO_2

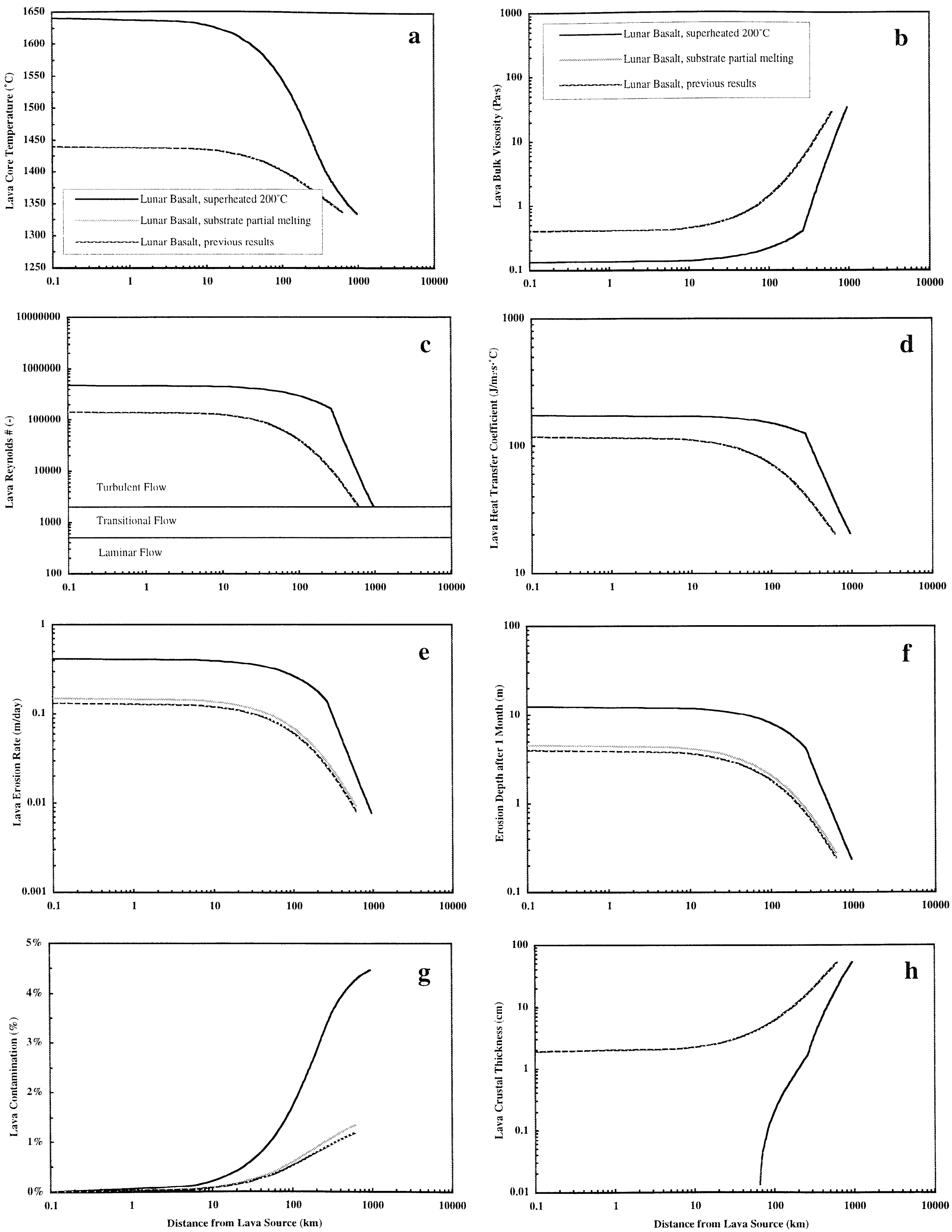

Figure 7. Modeling results for the emplacement of low-viscosity, lunar basalt lava flows over a lunar basalt substrate on the Moon. See also Table 5.

Table 5. Model Results for the Emplacement of Lunar Basalt Lava Flows Over Lunar Basalt Substrate on the Moon

Property	Nonsuperheated, No Partial Melting	Partial Melting of Substrate	Higher Flow Rate	Superheated Lava
Initial flow thickness, m	10	10	50	10
Maximum flow distance,* km	620	615	>9000	965
Maximum flow thickness,* m	14.8	14.8	>75	16.3
Maximum Reynolds number†	1.4E+05	1.4E+05	2.1E+06	4.7E+05
Maximum Prandtl number†	1.3E+03	1.3E+03	1.3E+03	5.7E+02
Maximum heat transfer coefficient,† J m^{-2} s^{-1} °C^{-1}	120	118	197	170
Maximum erosion rate,† cm d^{-1}	13	15	22	41
Maximum erosion depth after 1 week,† m	0.93	1.1	1.6	2.9
Maximum contamination,* %	1.2	1.4	>1.4	4.5
Maximum crustal thickness,* cm	52	52	>61	53

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

contents of lunar lavas result in higher-temperature, higher-density, lower-viscosity, more turbulent flows that have the potential to travel farther than terrestrial lavas (all else equal).

Regarding thermal erosion, our results predict low erosion rates (~10 cm d^{-1}) from thin (initially 10 m thick) lunar basalt lava flows over a lunar basalt substrate, such that long eruption durations (approximately months to years) are required to incise deep (tens to one hundred meters) channels. A model of substrate partial melting as suggested by *Hulme* [1973] only slightly increases thermal erosion rates. Other factors such as higher flow rates (thicker flows) or lava superheating could produce higher erosion rates over shorter eruption durations. In particular, a superheated lunar lava not only would have had a higher erosion rate (~40 cm d^{-1}), but would have traveled several tens of kilometers as an uncrusted channelized flow, consistent with the morphologies of most sinuous rilles.

Recent work by *Kauahikaua et al.* [1998] suggests that moderate-duration (months) eruptions of tube-fed basaltic lavas in Hawaii can erode considerable (>20 m) basaltic substrate. Thus we think that for long-duration (months to years) eruptions of low-viscosity lunar lavas, thermal erosion is a viable mechanism for the genesis of at least some sinuous rilles. If the implications of the work of *Wieczorek et al.* [2000] hold up, that superheated lavas erupted on the lunar nearside, then it seems even more likely that thermal erosion played an important role in the formation of the sinuous rilles. Much more work needs to be done to better understand the detailed dynamics of thermal erosion, and we are currently developing a more rigorous finite-volume model to investigate the thermal and physical conditions required for thermal erosion of the lunar surface.

Notation

c_{eff}	lava effective specific heat, J kg^{-1} °C^{-1}.
c_{gas}	gas specific heat, J kg^{-1} °C^{-1}.
c_l	lava liquid specific heat, J kg^{-1} °C^{-1}.
c_g	substrate specific heat, J kg^{-1} °C^{-1}.
E_{mg}	energy required to melt substrate, J m^{-3}.
ε	lava emissivity.
f_L	fraction of heat of fusion to partially melt substrate.
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^{-1} °C^{-1}.
k_c	lava crust thermal conductivity, J m^{-1} s^{-1} °C^{-1}.
k_{eff}	lava effective thermal conductivity, J m^{-1} s^{-1} °C^{-1}.
k_l	lava thermal conductivity, J m^{-1} s^{-1} °C^{-1}.
L_g	substrate heat of fusion, J kg^{-1}.
L_l	lava heat of fusion/crystallization, J kg^{-1}.
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.
μ_b	lava bulk viscosity, Pa s^{-1}.
μ_{eff}	lava effective viscosity, Pa s^{-1}.
μ_g	substrate melt viscosity, Pa s^{-1}.
μ_l	lava liquid viscosity, Pa s^{-1}.
ψ	slope of the ground, degrees.
Pr	lava Prandtl number.
ρ_b	lava bulk density, kg m^{-3}.
ρ_{eff}	lava effective density, kg m^{-3}.
ρ_g	substrate density, kg m^{-3}.
ρ_{gas}	gas density, kg m^{-3}.
ρ_l	lava density, kg m^{-3}.
Q_0	initial flow rate, m^2 s^{-1}.
$Q(x)$	flow rate, m^2 s^{-1}.
Re	lava Reynolds number.
σ	Stefan-Boltzmann radiative constant, J m^{-2} s^{-1} °C^{-1}.
$S(x)$	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^{-1}.
u_m	erosion rate of the substrate, m s^{-1}.
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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