9

Volcanic Vestiges

Pulling it Together

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9.1. INTRODUCTION

Commonly, students of volcanology study the effect large volcanic eruptions have on the local and global environment—as well they should, because it is within this arena that society and science meet to determine risk assessment and volcanic hazard mitigation. However, investigating the reverse is equally valuable and intriguing. By constraining how the environment effects the final morphology of volcanic deposits, we can quantitatively use volcanic geomorphology to reveal something about the environment in which the deposits were emplaced. In this way, volcanoes provide information about climate change on Earth (indicating the past location of a lacustrine shoreline, for example, or the previous extent of an alpine glacier; see Chapters 3 and 5) as well as the past and present conditions on other planets.

If a person's eyes can be thought of as windows into the soul, then a volcano can be seen as a window into a planet's deep interior. Until we are technologically capable of drilling to the center of the Earth or any other planet, volcanoes remain the single most important clue to the thermal, physical, and chemical behavior of the interior of a planet. For the Earth and the Moon, we have hand-sample analyses to give us direct information of lava compositions on those planetary bodies (see Chapters 2 and 6); and analyses of meteorites from Mars that have landed on Earth provide limited information of lava compositions there (see Chapter 4), although we cannot pinpoint the spot on Mars from whence these meteorites originated. Therefore, for the vast majority of solid bodies in the solar system, we must infer the lava and magma composition, as well as eruption and emplacement parameters (e.g., rheology, effusion rate, eruption duration), from the resulting volcanic morphologies. To interpret

Environmental Effects on Volcanic Eruptions: From Deep Oceans to Deep Space.
Edited by Zimbelman and Gregg, Kluwer Academic/Plenum Publishers, New York, 2000.

these morphologies accurately, however, "planetary volcanologists" must be able to decipher and remove the environmental effects, much the way a remote sensor analyzing the surface composition of a planet must remove the effects of the atmosphere on the spectra.

Because active eruptions have only been observed on Earth and Io, analytical and numerical models are used to predict the behavior and interpret the morphologies of extraterrestrial and deep-sea volcanic deposits, as well as for those terrestrial, land-based volcanoes that were not observed while active. In this chapter, we provide the reader with some basic numerical models obtained from the literature, as well as information on the thermophysical properties of various lavas and different environments likely to be encountered within the solar system.

9.2. INFORMATION

Most extant numerical models require that the user input various parameters (e.g., surface pressure, gravity, eruption temperature) to solve for the desired unknown (such as effusion rate, viscosity, or yield strength). In this section we include basic information on the thermophysical properties of a range of lava types, as well as the ambient conditions encountered in the solar system.

Table 9.1 lists typical parameters for lava types that have been proposed to exist on the terrestrial planets. Table 9.2 lists the thermophysical properties of the various bodies in the solar system that have been discussed in this book.

9.3. MODELS

Several numerical and analytical models designed to predict or interpret volcanic morphologies have been presented in the literature, and a thorough review would fill a book by itself! Here, we have selected a few of these models for the reader to experiment with. There is no room to document fully each model, so we strongly encourage interested students to go to the original references and investigate the assumptions that are an intimate part of any model. Natural volcanic systems are inherently complex, and currently no models can accurately predict the behavior of every aspect of a volcanic eruption, so almost by

Table 9.1. Thermophysical Properties of Common Lava Types

Eruption temperature (K)	Glass transition temperature (K)	Thermal diffusivity ($m^2\ s^{-1}$)	Eruption viscosity (Pa s)	Unvesiculated density[a] ($kg\ m^{-3}$)	Heat capacity ($J\ kg^{-1}\ K^{-1}$)	Associated composition
1425	1000	5.0×10^{-7}	10^2–10^5	2900	1200	Basalt[b]
1300	900	3.0×10^{-7}	10^5–10^7	2600	1125	Andesite[c]
1200	850	2.0×10^{-7}	10^7–10^9	2500	1050	Dacite[d]
1100	800	1.4×10^{-6}	$\geq 10^9$	2400	1000	Rhyolite[e]

[a] The actual erupted density may be a factor of 2 or 3 lower than listed here. Densities listed here assume no vesiculation. Typical densities for Hawaiian basalts are 1200–1500 $kg\ m^{-3}$ (Keszthelyi and Self, 1998). After Gregg and Fink (1996).
[b] Midocean ridge basalt (Griffiths and Fink, 1992a).
[c] Mount Hood andesite (Murase and McBirney, 1973).
[d] From Anderson and Fink (1992).
[e] Newberry rhyolite (Murase and McBirney, 1973).

Table 9.2. Thermophysical Properties of Environments Encountered in the Solar System[a]

Environment	Temperature (K)	Thermal expansion (K^{-1})	Thermal diffusivity ($m^2\ s^{-1}$)	Kinematic viscosity ($m^2\ s^{-1}$)	Density ($kg\ m^{-3}$)
Moon	4	—	—	—	—
Mercury	440	—	—	—	
Io	120	—	—	—	—
Mars	200	5.0×10^{-3}	7.2×10^{-4}	6.3×10^{-5}	0.2
Venus	730	1.3×10^{-3}	7.3×10^{-7}	4.5×10^{-7}	62.5
Earth (subaerial)	300	3.4×10^{-3}	2.3×10^{-5}	1.6×10^{-5}	1.2
Earth (submarine)	275	1.5×10^{-4}	1.0×10^{-7}	1.0×10^{-6}	1000

[a]After Griffiths and Fink (1992b) and Gregg and Fink (1996).

definition, a numerical model seeks to determine which parameters of the volcanic system are dominant. In what first appears to be a case of severe circular reasoning, however, each model must begin by making reasonable assumptions about which parameters can be neglected to simplify the system and make it tenable.

In this somewhat historical review, we begin by presenting models for lava flow emplacement, followed by the generation of explosive eruptions, Plinian eruption columns, and pyroclastic flows. In each case, the interested reader is strongly encouraged to find the original papers.

9.3.1. Lava Flow Emplacement

Modeling the emplacement of lava flows has advanced considerably in the past 25 years or so, but because lavas are complex mixtures of liquids, solids, and gases, it may well be another 25 years before a "universal" model of lava behavior can be adequately derived. Typically, the terrestrial volcanologist, investigating the emplacement of active flows, uses these models to determine properties that are the most difficult to measure in the field—commonly, lava viscosity. Similarly, the extraterrestrial volcanologist has only the solidified lava flow, and, most commonly, wishes to be able to determine the lava composition based on easily measured parameters (such as flow length, width, and thickness). No extant model yields lava composition, but many give yield strength and/or lava viscosity, and much research has been devoted to relating these properties to composition.

Nichols (1939) introduced Jeffrey's equation, given by

$$\eta = \frac{g\rho \sin\theta d^2}{nu} \tag{1}$$

where η is lava viscosity, g is gravity, ρ is lava density, θ is underlying slope, d is flow thickness, u is lava flow velocity, and n is an empirical constant that equals 3 for a broad flow and 4 for narrow flows. This model assumes a Newtonian rheology, and, therefore, is unlikely to be appropriate for lavas with a high yield strength, such as dacites and rhyolites (see Williams and McBirney, 1979, and Cas and Wright, 1987, for a more general discussion of lava rheology). This model has long been used to determine lava viscosities for extraterrestrial flows, assuming a "reasonable" flow velocity and density. At the simplest level, the problem with this application is twofold, however. First, reasonable flow velocities usually reflect observations made for Hawaiian lavas; it is important to note that Hawaiian volcanism is but one of many styles on Earth and may not be "typical" when compared with volcanic activity on other planets. Second, it is known that lava flow velocity varies with time (reflecting changes at the vent, or local changes in flow geometry, for example) and along the length and width of a flow. Although Jeffrey's equation may do an adequate job of estimating viscosity for active flows (where flow velocity, underling slope, thickness, and density can be accurately measured), but it can only produce estimates as accurate as are the assumptions for lava properties in the extraterrestrial context.

Hulme (1974) assumed a Bingham rheology to explain the formation of leveed lava flows on Earth and the Moon. He used the following relations to constrain the yield strength of leveed flows:

$$\Phi = \tfrac{2}{15} W^{2.5} - \tfrac{1}{4} W^2 + \tfrac{1}{6} W - \tfrac{1}{20} \tag{2a}$$

where Φ is a dimensionless quantity relating flow rate to liquid properties and external forces, and W is given by

$$W = \frac{w}{2w_b} \tag{2b}$$

where w is the half-width of a channelized flow and w_b is the width of the stationary levee. The quantity Φ can be related to yield strength through the relation

$$\Phi = Q\eta(g\rho)^3 \left(\frac{\theta}{\tau_y}\right)^4 \tag{2c}$$

where Q is volumetric flow rate, τ_y is lava yield strength, and other variables are as previously defined. Thus, by measuring the channel width and levee widths, and making "reasonble assumptions" about effusion rate, density, and lava viscosity, a yield strength could be obtained. Probably the single most difficult assumption to prove here is that the observed levees are simple levees, and were not formed through repeated lava overflows, or through accretion of rubble along the flow margin (cf. Sparks *et al.*, 1976).

Zimbelman (1985) used this model, and others also assuming a Bingham rheology, to constrain the yield strength of lavas on Ascraeus Mons, Mars. The relations he used are (Moore *et al.*, 1978)

$$\tau_y = \rho g d \sin\theta \tag{3a}$$

$$\tau_y = \frac{\rho g d^2}{2w} \tag{3b}$$

$$\tau_y = \rho g (2w - w_b) \sin^2\theta \tag{3c}$$

where d is flow thickness. Although it is clear that these relations provide the user with "a number," it is less clear precisely what this number reflects. Recent work by Peitersen and Crown (1998) shows that flow width ($2w$) varies dramatically along the length of a single basalt flow. It may be that these relations give some sort of "average" value for the yield strength of the entire flow. In other words, by using the above relations to compare different lava flows, you may be able to say something about the relative behavior of those two flows—but it is not obvious exactly what that is. Additionally, the reader should note the incompatibility between these three equations (leading to absurd equalities for some parameters). This is a direct consequence of the assumptions involved in the derivation of each equation, which vary drastically. Again, the utility of the results are only as good as the assumptions of each derivation.

It is important to note that these relations implicitly rely on a dimensionless parameter called the Graetz number, which is given by

$$\mathrm{Gz} = \frac{Qd}{Dxw} \tag{4}$$

where d is flow depth, D is lava thermal diffusivity, and x is flow length. Observations of Hawaiian flows indicate that most basalt lavas there stop advancing when Gz = 300, which is roughly the time during which the solidified surface crust of a flow has a thickness of approximately $1/3d$ (e.g., Pinkerton and Wilson, 1994). However, it is unlikely that this relation can be blindly applied to lavas of other compositions in other environments, because this limiting number (Gz = 300) is an empirical relation derived from Hawaiian lava flows.

Cooling rate clearly plays an important role in the final morphology of lava flows: The presence of lava pillows, generated only in subaqueous conditions, attests to that (see Chapter 5 and Gregg and Fink, 1995). Many workers have attempted to model the cooling rate of lava flows to better predict and interpret downstream changes in lava rheology and flow length (e.g., Crisp and Baloga, 1990; Pinkerton and Wilson, 1994; Fink and Griffiths, 1990). Other researchers have been concentrating on constraining the cooling rate of lava flowing within a lava tube (e.g., Sakimoto and Zuber, 1998; Keszthelyi and Self, 1998). Intuitively, the faster a lava flow cools, the shorter and thicker the resulting flow should be. However, a rapidly cooled lava flow may actually travel farther than a slowly cooled flow because the presence of a solid surface crust insulates the molten core of a flow from additional heat loss. Gregg and Fornari (1998) show that, for identical lava flows, basalts emplaced on the seafloor will travel approximately 30% farther than those on Earth's surface because of the enhanced cooling on the seafloor.

The rate of cooling of a lava flow is controlled by the ambient conditions, the lava eruption temperature, viscosity, and velocity (a proxy for the rate of heat advection within a flow) (e.g., Crisp and Baloga, 1990; Fink and Griffiths, 1990). For subaerial terrestrial flows, Martian flows, as well as lunar, Mercurian, and Ionian lavas, the primary cooling mechanism is radiative cooling. The equation for heat flux from a radiating lava flow is given by

$$F_r = \varepsilon\sigma\{(1-f)(T^4 - T_a^4) + f(T_{sc}^4 - T_a^4)\} \tag{5}$$

where F_r is radiative heat flux from the flow; ε is lava emissivity; σ is the Stephan–Boltzmann radiative constant; f is the fraction of the flow covered with a solidified crust; T is temperature, and subscripts "a" and "sc" indicate "ambient" and "surface crust". Without even plugging in the numbers, it is evident that for basaltic lavas on Earth, the ambient temperature can be safely neglected, but that on Venus, it might be worthy of consideration.

For lavas emplaced in sufficiently thick atmospheres (such as the deep seafloor or the surface of Venus), convective cooling plays an important role. The heat flux from a connectively cooled body is given by (Fink and Griffiths, 1990; Gregg and Greeley, 1993)

$$F_c = \rho_a c_a \gamma\left(\frac{\alpha_a g D_a^2}{\nu_a}\right)\{(1-f)(T - T_a^{4/3})^{1/3} + f(T_{sc} - T_a)^{4/3}\} \tag{6}$$

where ρ is lava density, c is heat capacity, γ is an empirical constant equal to 0.1 (Turner, 1973), α is the coefficient of thermal expansion, g is gravitational acceleration, D is thermal diffusivity, ν is kinematic viscosity, and other variables are as previously defined.

Griffiths and Fink (1992a) show that for lava flows on the seafloor, convective cooling dominates throughout lava flow emplacement. On land, radiative cooling dominates until the surface temperature of the lava flow $\leq 250°C$ (significantly lower than the solidification temperature of silicate lavas; Table 9.1). Convective cooling is the dominant process on Venus until the lava surface temperature falls below $\sim 750°C$, when radiative cooling begins to take over.

Once a surface crust has formed over a lava flow, the molten interior of the flow cools by thermal diffusion through the overlying crust (Crisp and Baloga, 1990). This is also the primary mechanism for cooling lava within lava tubes (e.g., Sakimoto and Zuber, 1998; Keszthelyi and Self, 1998). Therefore, although the formation of lava tubes may be preferentially enhanced on Venus and the seafloor (see Chapter 5) in comparison with other volcanic environments, once a tube forms, the behavior of the lava varies little in different ambient conditions. Gravity helps to control the velocity at which lava is capable of flowing within the tube, and also controls the width of an unsupported lava tube roof once the lava had drained away (Oberbeck *et al.*, 1969). Thermally, however, lava tubes should behave similarly under various ambient conditions.

9.3.2. Explosive Eruptions

Pyroclasts can be generated by the violent depressurization of magmatic volatiles, or by the interaction of hot magma with near-surface ground water, ground ice, or shallow lakes, ponds, or seas. On Earth, H_2O and CO_2 are the most common magmatic volatiles (e.g., Cas and Wright, 1987; Sparks *et al.*, 1997); although CO may dominate on the Moon (see Chapter 6) and CO_2 may prevail on Venus and Mars (see Chapters 4 and 5).

A combination of cooling rate, effusion rate, and lava viscosity may exert the strongest controls on the final morphology of effusive lava flows (Fink and Griffiths, 1990), although only one of these—cooling rate—is determined by the ambient conditions. The production and distribution of pyroclastic deposits, however, is closely tied to the atmospheric pressure and temperature, as well as the intrinsic magmatic properties such as volatile content, viscosity, and temperature. The surface pressures of the different environments examined in this book vary widely—from the vacuum of space to the intense pressure (>250 MPa) on the deep seafloor.

Models for the generation and emplacement of pyroclastic deposits have focused on understanding the generation and subsequent collapse of Plinian eruption columns (e.g., Sparks, 1978; Sparks *et al.*, 1997) because these types of eruptions are historically most closely associated with loss of life and property damage (e.g., Tilling, 1989). Wilson *et al.* (1978) showed that the maximum height of a Plinian eruption column can be predicted from

$$H_p = 8.2R^{1/4} \tag{7a}$$

where R is the steady-state energy release in watts and is given by

$$R = \rho v \pi r^2 c(T - T_a)E \tag{7b}$$

where ρ is the bulk density of the erupting fluid (a mixture of solids, liquid, and vapor), v is velocity, r is the vent radius, c is specific heat, T is temperature, and E is an "efficiency factor" that measures how efficiently heat is converted to potential or kinetic energy. Results from recent modeling show that the type of gas has a strong influence on the eruption column height. For example, Campbell *et al.* (1998) demonstrated that it would be very difficult to sustain a CO_2-dominated eruption column on Venus—the high density of the column would cause it to readily collapse (see Chapter 5). In contrast, SO and SO_2 have relatively low densities, contributing to the large eruption plumes observed on Io.

Additionally, the lower the surface pressure, the greater the amount of gas is able to exsolve from the magma by the time the magma reaches the surface—which should result in a greater range for the pyroclasts (cf. Wilson and Head, 1981). A thin or nonexistent atmosphere will also exert less drag on any ejected magma fragments, resulting in a greater dispersal for given eruption conditions. The range, X, of pyroclasts can be obtained as (Wilson and Head, 1981)

$$X = r - \frac{v}{g}\left(\frac{dr}{dh}\right)\left[v + \sqrt{v^2 - 2gh}\right] \tag{8}$$

where X is the range (distance from the vent), r is vent radius, v is rise velocity, and (dr/dh) is the slope of the vent wall at the fragmentation level at depth h in the conduit. Clearly, for a given set of eruption conditions, planets with gravity lower than Earth's will have more widely dispersed pyroclastic deposits (see Chapters 4, 6, and 7).

9.4. MAGMA INTRUSION

The means by which magma reaches the surface is a vital part of the volcanic system. For example, near-surface dikes have created spectacular landforms on Venus (see Chapter 5), and apparently created some linear rilles on the Moon (see Chapter 6). The formation of dikes and of magma storage systems within or beneath the crust (see Chapter 5) is apparently more closely controlled by density differences between the host rock and the magma and by the pressurization within the magma storage system, than by any "atmospheric" or superficial parameters (e.g., Wilson and Head, 1981; Ryan, 1994, and references therein). Although gravity weakly enters into most models designed to predict the ascent rate of magma within a dike of given dimensions, it is overshadowed by density differences and regional stresses (e.g., Wilson and Head, 1981; Ryan, 1994). And while a basaltic dike rising through hydrothermally cooled oceanic crust may cool more rapidly than an identical dike ascending through the hot Venusian crust, again, controls other than temperature (e.g., lava viscosity, velocity, and relative density) are more important.

9.5. SUMMARY AND CONCLUSION

Historic eruptions have clearly revealed how volcanism can strongly impact local and global climate, and much research has been devoted to predicting and understanding this phenomenon. For those interested in deciphering the volcanic clues left enticingly on the surfaces of other planets, however, investigating the converse relation—how the local and global environment effects eruption dynamics—is essential. Volcanoes have been identified on solid bodies throughout the solar system, but Earth remains the only place where actively flowing lava has been clearly observed, and Io is the only other place in the solar system where active volcanism has been witnessed. Thus, it is imperative that we learn to interpret extraterrestrial and submarine volcanic morphologies by first quantifying and removing the specific environmental effects.

Within this book, we have presented volcanism in the cold vacuum of space (e.g., Earth's Moon and Io; see Chapters 6 and 7), the wispy-thin atmosphere of Mars (Chapter 4), beneath tens to hundreds of meters of ice (Chapter 3), at the dark, cold bottom of Earth's oceans and on the scorching surface of Venus (Chapter 5). These chapters introduce an astonishing array of volcanic morphologies and eruptive styles, and when one considers the manifestations of "exotic" and "ice" lavas (Chapter 8), the morphologic possibilities are virtually endless.

We have tried to demonstrate that the kind of environment a volcano erupts in is just as important as many of the more commonly studied variables, such as lava composition and rheology. The models presented in this chapter are included so that the interested readers can insert the appropriate values to see for themselves the role ambient conditions play in the emplacement of volcanic deposits. Although this book contains a wealth of information on the environmental effects on volcanic eruptions, there is still a vast amount of research to be done. Until we, as volcanologists, are able to do thorough fieldwork on the surfaces of other planets, we will never know for sure if our models and interpretations are correct. And until then, we will continue to observe, model, and test—only to be required to observe everything again from a new perspective generated by ongoing analyses and missions.

9.6. REFERENCES

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