ENVIRONMENTAL EFFECTS ON VOLCANIC ERUPITONS: FROM DEEP OCEANS TO DEEP SPACE

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CHAPTER 3. VOLCANISM AND ICE INTERACTIONS ON EARTH AND MARS PAGES 39-79.

by Mary G. Chapman,[1] Carlton C. Allen,[2] Magnus T. Gudmundsson,[3] **Virginia C. Gulick**,[4] Sveinn P. Jakobsson,[5] Baerbel K. Lucchitta,[1] Ian P. Skilling,[6] and Richard B. Waitt[7]

[1]US Geological Survey
2255 N. Gemini Drive
Flagstaff, AZ 86001, USA

[2]Lockheed Martin Space Operations Co.
Mail code C23
2400 NASA Road 1
Houston, TX 77058, USA

[3]Science Institute, University of Iceland
Hofsvallagata 53
107 Reykjavik
Iceland

[4]NASA Ames Research Center
MS 245-3
Moffett Field, CA 94035, USA

[5]Icelandic Institute of Natural History
Hlemmur 3
IS-105 Reykjavik, Iceland

[6]Dept of Geology
Rhodes University, PO Box 94
6140 Grahamstown
South Africa

[7]US Geological Survey
Cascades Volcano Observatory
5400 MacArthur Blvd.
Vancouver, WA 98661, USA

Corresponding author: Mary Chapman[1], 520-556-7182; Fax: 520-556-7014, mchapman@flagmail.wr.usgs.gov

3.1. INTRODUCTION: VOLCANO/ICE INTERACTIONS ON EARTH AND MARS

When volcanoes and ice interact, many unique types of eruptions and geomorphic features result. Volcanism appears to occur on all planetary bodies, but of the inner solar system planets, ice is limited to Earth and Mars. Earth, the water planet, is covered by ice wherever the temperature is cold enough to freeze water for extended periods of time. Ice is found in sheets covering the Antarctic continent and Greenland, as glacial caps in high mountainous regions, as glaciers in polar and temperate regions, and as sea ice in the northern and southern oceans. With changes in climate, land-mass, solar radiation, and Earth orbit, ice masses can contract or expand over great distances, as occurred during the Pleistocene Ice age. The Earth is still in an ice age--at the beginning of the Cenozoic, 65 million years ago, our planet was ice-free. In fact, there is now so much ice that about 70% of the worlds total fresh water is contained within the Antarctic ice sheet.

It is less well known that Mars also has a significant amount of ice, including polar caps somewhat like (those on) Earth. Its north polar cap consists of mostly water ice, its south polar cap is surfaced by CO_2 ice [*Kieffer*, 1979]. The most important reservoir of ice, however, is retained in the ground as pore fillings or segregated masses. The average mean temperature on Mars is -80° C overall, -60° at the equator [*Fanale et al.*, 1992], resulting in a permafrost layer (ice rich?) ranging from about 1 to 2 km thick at the equator to 3 to 6 km at the poles [*Rossbacher and Judson*, 1981; *Clifford*, 1993]; a common estimate for midlatitude permafrost thickness is 2 km [Squyres et al., 1992]. However, in near-surface rocks or soil between about 50 degrees north and south latitude [*Mellon and Jakosky*, 1995], ice may not be retained, because in the martian thin atmosphere water vapor migrates out of the ground in the warmer equatorial areas. On the other hand, complexities of the surface expressed in varying albedo (relative brightness), thermal inertia (ability to absorb and release heat), and orbital obliquity (angle between equatorial plane and orbit) may allow for the existence of near-surface ice even in this region [*Mellon and Jakosky*, 1995, *Paige*, 1992]. Geological observations attest to a large reservoir of ground ice, near the surface or at depth, almost everywhere on Mars, leading Carr [1995] to suggest an ancient water layer several hundred meters thick (if evenly distributed). These observations include flow lobes on crater ejecta, features possibly attributed to thermokarst such as chaotically collapsed ground and irregular depressions, flow lobes on scarps that resemble rock glaciers, and numerous fluvial valleys and outflow channels. Water from channels appears to have pooled in the northern lowlands, forming a lake or ocean

[*Witbeck and Underwood*, 1983; *Lucchitta et al.*, 1986; *Parker et al.*, 1989, 1993; *Jöns*, 1990; *Baker et al.*, 1991; *Scott et al.*, 1992; *Lucchitta*, 1993]. This ocean, though blanketed by dust, may remain as a vast reservoir of ice.

Our knowledge of planetary volcano and ice interactions begins on Earth. Many terrestrial stratovolcanoes have considerable ice covers on their slopes and some volcanic regions are covered by large ice sheets and caps. These different settings give rise to two types of eruptions: (1) those in alpine situations beneath mountain snow or summit and valley glaciers and (2) those beneath broad continental-scale glaciers or ice sheets. Another type (3) are not true eruptions but result from explosions caused by lava flowing over water/ice-saturated ground and form phreatic craters. Terrestrial volcano/ice interactions of eruption types 1 and 2 both involve large volumes of meltwater and therefore usually generate floods, ranging from water floods to mudflows or lahars (an Indonesian term) composed chiefly of volcaniclastic materials. Jökulhlaup is an Icelandic term for glacier outburst flood.

On Earth, sub-ice sheet eruptions (type 2) from fissures or point sources produce pillow lavas and hyaloclastites (vitroclastic tephra produced by interaction of water and lava). Overlain by meltwater and ice, such eruptions from fissure sources form hyaloclastic ridges, while those from point sources form mounds. If the eruption penetrates the ice, subaerially extruded layered lava may cap these volcanoes, resulting in a tuya. This feature is commonly and perhaps incorrectly called a table mountain, since a table mountain in a geomorphic sense refers to a flat-topped mountain of any origin and translates directly to mesa.

Volcanic accumulations of hyaloclastites, combined with pillow lavas, irregular intrusions and occasionally subaerial lavas, are widely exposed in the volcanic zones of Iceland. Moberg, a generic Icelandic term, literally means "rock that looks like peat" and is used to describe all brownish, rich in palagonite (hydrated volcanic glass), basaltic, hyaloclastic rocks of Iceland, including all subglacial volcanic materials that formed during the upper Pleistocene but also during Recent times. Therefore the use of the term moberg ridge, when discussing a hyaloclastic ridge, is a misnomer.

The origin of the hyaloclastite volcanoes was for a considerable time a matter of dispute. Peacock [1926] was first to present a "volcano-glacial" origin for hyaloclastites. Noe-Nygaard [1940] gave a description of a subglacial volcanic eruption in Iceland. Independently in the 1940s, Kjartansson [1943] in Iceland and Mathews [1947] in British Columbia presented the hypothesis that the Pleistocene hyaloclastite mountains are piles formed in subglacial eruptions. The work of Van Bemmelen and

Rutten [1955] in the northern part of Iceland established the theory. Moore & Calk [1991] provided definitive geochemical studies of Icelandic tuyas. Because of the late acknowledgment of subglacial volcanism as a process, it is a relatively young science and geologists are unfamiliar with the subject. This chapter seeks to help alleviate this problem with detailed terrestrial examples of recent volcano/ice eruptions, lithologic details of hyaloclastic ridges and a tuya deposit, and descriptions of recent lahars and jökulhlaups.

Our planetary discussion of volcano/ice interactions is limited to Mars, whose surface contains ice and a variety of volcanic landforms, including volcanic plains, ancient highland volcanoes, and two large volcanic regions, Tharsis and Elysium (see chapter 4). Crater density data suggest that volcanism continued throughout Mars' geologic history [*Tanaka et al.*, 1992]. Because extensive morphological evidence exists for ground water and ground ice throughout the planet's history, ground-ice and frozen waters likely interacted with sub-ice volcanoes. In fact, there are martian geomorphic features that have been interpreted to be analogous to those terrestrial edifices that form from all three types of sub-ice eruptions. Therefore, in addition to the discussion of terrestrial volcano/ice interactions, other sections in this chapter note these examples of martian analogs.

Volcano/ice interactions on Earth produce catastrophic lahars and distinctive edifices. On Mars these types of interactions seem to produce geomorphic change on two timescales. The emplacement of the volcano itself can produce large scale geomorphic signatures (cones, tuyas, etc). On a more gradual timescale the subsurface heat source could heat and cycle ground water to the near surface environment over periods of thousands to millions of years, depending upon the size of the intrusion. The resulting hydrothermal systems may also provide hospitable environments for the evolution of life on Mars. Therefore, the important implications of groundwater and snow/ice perturbations from hydrothermal systems will also be discussed in this chapter.

3.2. Nevado del Ruiz, Colombia, November 1985--eruption beneath an alpine ice cap

Although, Baker et al [1991] have suggested the possibility of local snowfall on elevated areas, surrounding the hypothetical paleo-ocean of Mars, certain volcano/ice interactions may be unique to the Earth, where precipitation occurs regularly. This limitation may be the case for eruptions beneath ice-capped volcanoes. A recent, relatively small example of this type of eruption occurred on November 13, 1985, at Nevado del Ruiz, Colombia. Nevado del Ruiz is 5400 m high, the

northernmost-active volcano of the Andes, and a typical subduction zone andesite [*Williams et al.*, 1986] volcano, that produced a horrible catastrophe, killing more that 23,000 people living 72 km away, in the city of Armero. The 1985 eruption is only the most recent event in a series of twelve eruptive stages that have occurred in the past 11,000 years at the 0.2 m.y. old Ruiz volcano [*Thoret, et al.*, 1990].

About twenty-five to thirty percent of the ice on the ice cap of Nevado del Ruiz was fractured and destabilized by earthquake and explosion; nearly 16% (4.2 km^2) of the surface area of the ice and snow (about 0.06 km^3 or 9% of the total ice and snow volume) was lost [*Thouret*, 1990]. Large volumes of meltwater were produced (20 million m^3 and combined peak discharge of 80,000 m^3/s) that flowed downslope, liquefied some of the new volcanic deposits, and generated avalanches of saturated snow, ice and rock debris within minutes of the eruption [*Pierson et al.*, 1990; *Pierson*, 1995]. Lahar extents varied from 7.2 to 102.6 km from the vent.

Thouret [1990] estimates that the meltwater volume released during the 1985 eruption was 2-4 times that which was incorporated into the lahars. This non-contributing water was: (1) included in snow avalanches or bound up in pyroclastic surge deposits, (2) incorporated in the phreatic explosive products, (3) sublimated into steam, or (4) stored within the current ice cap [*Thouret*, 1990].

The key lesson of the Nevado del Ruiz eruption are: (1) catastrophic lahars (section 3.7) can be generated on ice- and snow-capped volcanoes by relatively small eruptions, (2) because processes acting predominantly at the surface mobilize lahars, the surface area of an ice cap can be more critical than total ice volume when considering lahar potential, (3) placement of hot rock debris on snow is insufficient to generate lahars--the two materials must be mechanically mixed together (in pyroclastic flows, surges, and mixed avalanches) for sufficiently rapid heat transfer, (4) lahars can increase their volumes significantly by entrainment of water and eroded sediment, and (5) valley-confined lahars can maintain relatively high velocities and can have catastrophic impacts as far as 100 km downstream [*Pierson et al.*, 1990]. Based on these lessons, one might expect much larger lahar volumes and extents on the comparatively larger volcanoes of Mars.

Many factors continue to render the Ruiz domes unstable, including deeply dissected troughs, strongly hydrothermally altered summit flanks, currently glaciated high cliffs, and steep unstable ice cap margins [*Thoret, et al.*, 1990].[24] The presence of a large volume of hydrothermally altered rock within the volcano not only increases the potential for edifice collapse because the altered rock is less competent, but also increases its potential to become a saturated, liquefied debris flow because the rock can hold more water [*Carrasco-Núñez et al.* 1993]. Alteration gases and waters are emitted from fumaroles and thermal springs on the flanks of the volcano and are likely to be derived from a two-phase vapor-brine envelope surrounding the magmatic system [*Giggenbach et al.*, 1990].

Hydrothermal alteration leading to unstable summit flanks may have also occurred on Mars. Olympus Mons and Apollinaris Patera are bound by steep scarps, indicating failure of summit flanks and Olympus Mons appears to be surrounded by some type of debris lobes (aureole; section 3.10.3). Furthermore, hydrothermal discharge from fumaroles and springs may have formed valleys on the flanks of martian volcanoes (section 3.10).

3.3. GJáLP, VATNAJÖKULL, OCTOBER 1996--ERUPTION BENEATH AN ICE SHEET

Because outflow channels may have accumulated water in ponded situations, subglacial eruptions under broad ice sheets may have been a more likely occurrence on Mars than eruptions under alpine glaciers. Subglacial eruptions are frequent in Iceland, and the majority occur within the Vatnajökull ice cap (fig. 3.1.A; [*Thorarinsson*, 1967, 1974; *Björnsson and Einarsson*, 1990; *Larsen et al.*, 1998]). However, owing to low activity in recent decades, the first such eruption to be monitored in any detail was the 13-day-long fissure eruption in Gjálp, Vatnajökull, in October 1996 [*Gudmundsson et al.*, 1997]. The progress of the eruption is shown schematically in figure 3.1.B. At the central part, the eruption melted its way through 500 m of ice in 30 hours, forming a subaerial crater (Figs. 3.2 and 3.3). The eruption produced basaltic andesite with SiO_2 content of 54% [*Gudmundsson et al.*, 1997]. A thin layer of tephra was dispersed over about 10,000 km^2 but its volume is no more that 1-2% of the total erupted. The eruption formed a 6-km-long ridge with three peaks in the bedrock (Fig. 3.1.B), with an estimated volume of 0.7 km^3. The bedrock relief increased by 200-350 m; the highest peak is about 150 m below the initial ice surface. The new ridge is partly built on a ridge formed in a similar eruption in 1938.

Data on melted volumes allowed calorimetry to be used to infer roughly the eruption rate with time and provide an estimate of the total volume erupted [*Gudmundsson et al.*, 1997]. Magma discharge was highest in the first four days (melting rate 0.5 km^3/day) and by the end of the eruption melted ice had produced 3 km^3 of water. After 100 days the melted volume was 4 km^3 and the total reached 4.7 km^3 in January 1998. The high rate of heat transfer during the eruption suggests that fragmentation of magma into glass (hydroclasts or vitriclasts) was the dominant form of activity, not eruption of pillow lava. About 20% of the melting during the eruption occurred over the subglacial path of the water, indicating that its temperature was 15-20°C when it left the eruption site. This high temperature must have greatly enhanced the opening of subglacial drainage channels.

The meltwater from the eruption accumulated in the Grímsvötn caldera lake for 5 weeks before it was released in a very swift jökulhlaup. An estimated volume of 3.5 km^3 was drained in two days, flooding the Skeidarársandur outwash plain to the south of Vatnajökull, destroying bridges and roads.

In June 1997, inspection of the exposed uppermost 40 m of the edifice showed that it was mainly made of clastic rocks with ash to lapilli sized glass particles, only a small fraction was crystalline rock (max. diameter of fragments 20 cm). The temperature at 0.5 m depth was 60-70°C. In June 1998, ice flow had covered the edifice. Monitoring of heat flow and other observations may show to what extent the ridge becomes palagonitized and consolidated.

Significant storage of meltwater at the eruption site did not occur at Gjálp, as explained by applying the theory of water flow under temperate glaciers [*Paterson*, 1994; *Björnsson*, 1998]. The direction of flow is down the slope of a potential function determined by (i) the slope of the bedrock, and (ii) the slope of the overlying ice surface. The ice surface slope is nine times more influential and controls the flow direction except in areas of very rugged bedrock (flow may be uphill, see Fig. 3.1.A). Thus, meltwater was drained away from the beginning, as if it were flowing on the ice surface. (Note the interesting implication for Mars, where overlying, non-polar surface ice has long since sublimated away and paths of meltwater are assumed to follow bedrock slopes.) After depressions had formed, subglacial drainage continued, since melting due to the heat of the water prevented closure of the subglacial conduits.

The opening formed in the ice (Fig. 3.3) remained only 200-300 m in diameter throughout the eruption; wall melting was apparently compensated by ice flow. The water level at the eruption site was 150-200 m below the original ice surface; the ice cap in the surrounding area is 600-700 m thick, hence, the height of the water column was 70-80% of the ice thickness. Outside the subsidence cauldrons, no signs of enhanced ice flow have been detected, indicating that for temperate glaciers the effects of subglacial eruptions on ice flow are localized.

3.4. SUBGLACIAL VOLCANIC HYALOCLASTIC RIDGES

Subglacial hyaloclastic ridges are unique indicators of volcano/ice interaction and appear to have only been described in Iceland. However, some formations on the volcanic island Jan Mayen in the Norwegian Sea [*Imsland*, 1978] may be classified as short ridges. Although unreported, hyaloclastite ridges probably occur in other volcanic subarctic or arctic regions.

A distinction is made between three main phases in the generation of a subglacial volcano: pillow lava, hyaloclastites, and subaerial lava. A pile of undegassed pillow lava forms if the hydrostatic pressure exceeds the gas pressure of the magma. Typically, only a minor amount of hyaloclastites form between pillows. Pillow lavas are often found at the base of hyaloclastite mounds, ridges and tuyas. The thickness of basal basaltic pillow lava piles often exceeds 60-80 meters and a 300-m-thick section has been reported [*Jones*, 1970].

The depth at which effusive eruption of pillow lava transitions to fragmentation of hyaloclastites is not well known. The transition depth appears to be controlled by two factors, (1) the volatile content of the magma, and (2) the hydrostatic pressure at the vent. Eruption rate may also be an influencing factor. On the basis of Late Pleistocence formations in Iceland, Jones [1970] suggested basaltic phreatomagmatic explosions produce hydroclastites at a water depth less than approximately 100-150 m. Recent field studies in Iceland, however, suggest that a somewhat greater depth may be typical for basalts [*Kokelaar*, 1986].

Hyaloclastites of Icelandic deposits can be subdivided into two main types. Type one is poorly bedded, unsorted, hyaloclastite [*Jakobsson & Moore*, 1982] that probably was quenched and rapidly accumulated below sea level without penetrating the surface. The coarse-grained, often massive core of many of the larger tuyas and ridges may have formed in this way. Type two is more fine grained than the lower, more massive part and forms at shallow water depths (approximately 20-30 m in the submarine Surtsey eruption; *Kjartansson*, 1966). At these shallow depths, the phreatic eruptions start to penetrate the water level, and bedded hyaloclastite (tephra) is deposited by air fall or base surge on ice or land and sediment gravity flows into ponded water (e.g. in the Surtsey and Gjálp eruptions). In some cases, hyaloclastites form flows, but their extent is not known and may be limited to southeast and south Iceland. Walker & Blake [1966] have described such an Early Pleistocene mass of hyaloclastite with associated pillows and columnar basalt in Dalsheidi in SE-Iceland. It is interpreted as a flow beneath a valley glacier in a channel previously excavated by englacial floods. It is ≤ 300 m thick and may originally have been 35 km long and about 0.5 km wide.

Lava caps the hyaloclastite section of the tuyas and possibly about 10-15 % of the Icelandic hyaloclastite ridges. This lava is comparable to other subaerial lava; sheet lava is most common but massive, simple flows also occur. The lava grades into foreset breccia and sometimes into gently dipping, degassed pillow lava. Besides regular feeder dykes, apophyses and small irregular intrusions are common.

Moore & Calk [1991] studied the geochemistry of four Icelandic tuyas and one hyaloclastite ridge. A systematical upward change in rock composition was found in most cases, exemplified by decrease of the MgO content. The upward decrease in sulfur content is thought to be a measure of the change in depth of water (or ice) cover at the vent during eruption.

Evidence for gravitational sliding [*Furnes & Fridleifsson*, 1979] and collapse in the walls of the subglacial volcanoes is commonly seen. Evidently the walls became unstable when the confining ice retracted. Prior to consolidation the hyaloclastites are easily eroded by glaciers and running water both during and after the eruption.

The bulk of the hyaloclastites are consolidated and altered. The volcanic glass is thermodynamically unstable and the basaltic glass alters easily to palagonite [*Fisher & Schmincke*, 1984]. Palagonite is a complex substance and variable in composition and the overall process of alteration is still imperfectly understood. Alteration primarily occurs under two different types of conditions, in short-lived, mild, hydrothermal systems and at weathering conditions [*Jakobsson*, 1978]. Intrusions in the hyaloclastite provide the heat for the hydrothermal system in the Surtsey tuya [*Jakobsson & Moore*, 1986]. The basal pillow lava pile possibly serves as a heat source in subglacial and subaquatic volcanoes [*Sigvaldason*, 1968]. The identification of bacteria in near-surface subaerially formed palagonite has led Thorseth et al. [1992] to suggest that microbes have played an important role in the development of certain textures in porous palagonite.

The distribution of Icelandic Late Pleistocene and Recent eruption units having significant hyaloclastite is shown in Figure 3.1.A. Most of these units belong to the Moberg Formation [*Kjartansson*, 1960], formed during the Late Pleistocene (0.01 - 0.78 Ma), and are predominantly of subglacial origin, though a smaller part is formed in lacustrine and marine environments. Tuff and tuff breccia are common along with pillow lava. Moraines and fluvioglacial deposits are often intercalated with the hyaloclastites. Intraglacial and interglacial lavas are also an important constituent. Subglacial eruptions within the ice caps of Vatnajökull and Myrdalsjökull have continued producing hyaloclastites up to the present.

Subglacial hyaloclastite ridges and mounds are distributed throughout the volcanic zones of Iceland (Fig. 3.1.A). The number of exposed units is not known but may easily exceed 1000-1200. A recent study (unpublished) in the northern part of the Western Volcanic Zones (Fig. 3.1.A) has identified 83 subglacial volcanoes, of which 59 are hyaloclastite ridges or mounds and 24 would be classified as tuyas. The most common lithofacies is poorly bedded tuff breccia, often with pods and lenses of pillow lava, and finely bedded tuff. Basal pillow lava is found in 30 mountains, 8 are solely made of pillows. The length of the ridges in this area varies between 1 and 18 km and the width between 0.5 and 2.2 km. Maximum height is commonly 200-400 m.

Two examples of typical medium sized subglacial hyaloclastite ridges occur in the Western Volcanic Zone (Fig. 3.4A, B). Pillow lava is seen at the base of both ridges. The distinction between the ridges and tuyas is not clear as demonstrated by the mountain Stora-Björnsfell (Fig. 3.4C). The basal pillow lava appears to have erupted from a fissure. As the eruption progressed, activity concentrated at one vent that eventually produced the subaerial lava cap. Nine ridges in the Western Zone have remnants of subaerial lava on the top; therefore, these units, like the tuyas, have penetrated the ice sheet.

Jones [1969, 1970] published a structural analysis of the subglacial ridge Kalfstindar in the Western Volcanic Zone. Kalfstindar (Fig 3.5) is not a typical ridge, it is unusually large, has an exceptionally thick basal pillow pile, and may be composed of a few smaller eruption units.

The area between Vatnajökull and Myrdalsjökull is dominated by hyaloclastite ridges (Fig. 3.6 and 3.17.A), no basaltic tuyas are found in the area. The ridges are larger than in the Western Zone, the longest eruption unit is reported to be 44 km long [*Vilmundardottir and Snorrason*, 1997]. The ridges in this area commonly have relief of 300-400 m. A number of ridges have been found with radio-echo soundings under the western part of Vatnajökull [*Björnsson*, 1988].

Silicic subglacial hyaloclastites are found within Pleistocene central volcanoes in Iceland, although comparatively rare [*Furnes et al.*, 1980]. Typical rhyolitic formations are ridges 2 to 2.5 km long and 0.7 to 1 km wide. Well-known examples are Bláhnúkur in the Eastern Volcanic Zone [*Saemundsson*, 1972] and Hlidarfjall in the Northern Zone [*Jonasson*, 1994]. A few examples of rhyolitic tuyas have been briefly described (e.g. Kirkjufell in the Eastern Zone [*Saemundsson*, 1972] and Höttur in the Western Zone [*Grönvold*, 1972]).

Why hyaloclastite ridges form in some cases and tuyas in other cases is unclear. Analyses of rocks from the northern half of the Western Volcanic Zone have showed that the distinction is not governed by magma chemistry. Tectonic control may be important since in an eruption on a typical rifting related fissure, the available magma will be quickly erupted along the length of the fissure. The hyaloclastite ridges most likely form in such eruptions. In contrast, the tuyas in Iceland appear to form in eruptions along short tectonic fissures producing large volumes of magma per unit length (Fig. 3.7). Possibly, the underlying tectonics in tuya-forming eruptions are vertical crustal movements due to loading/unloading when overlying ice sheet advanced/retreated in response to climatic fluctuations.

The height of a tuya/ridge subaerial lava cap must have some relation to the ice thickness, but the relationship may not be simple. The level of a meltwater lake determines the height where subaerial flow of lava starts. In temperate ice caps the water level of subglacial lakes rarely rises to 90% of the ice thickness, the level required to float the ice. Such lakes are drained in jökulhlaups at levels several tens of meters below the floating level [*Björnsson*, 1988]. Thus, within temperate glaciers, semi-stable water-levels higher than 80% of the ice thickness appear unlikely (see section 3.4). Cold based glaciers are frozen to their beds, and if water is to flow subglacially, brittle failure of the ice-bedrock contact must take place. However, due to the strength of the ice-bedrock contact, water flow along the ice surface may be dominant in eruptions within thin cold-based glaciers. No observations exist, but in this case, the water level of an meltwater lake may be similar to the ice thickness.

At the peak of the last glaciation in Iceland, the ice sheet would have had a mean thickness of about 1.0 km, double the typical height of ridges and tuyas. This may suggest that the tuyas and lava-capped ridges were formed during intervals with less extensive glacial cover.

3.5. SUBGLACIAL VOLCANIC TUYAS

Hyaloclastic ridges capped by subaerial lavas are called tuyas. Basaltic hydrovolcanic sequences with tuya or table-mountain form, which have been interpreted as the products of subglacial and englacial eruptions, have been recorded at several localities on Earth. All of the localities are in present day high-latitude areas, reflecting the low preservation potential of such sequences and/or problems of interpretation in areas situated far from present day glaciers or ice-sheets. Sequences have been recorded from Iceland [*Jones*, 1969, 1970; *Saemundsson*, 1967; *Sigvaldason*, 1968; *Allen et al.*, 1982] British Columbia [*Mathews*, 1947; *Allen et al.*, 1982; *Hickson and Souther*, 1984] and Alaska [*Hoare and Coonrad*, 1978].

Brown Bluff volcano [*Smellie and Skilling*, 1994; *Skilling*, 1994] is a 1Ma-old [*Rex*, 1976] tuya, located at the northern tip of the Antarctic

Peninsula (Fig. 3.8). A number of factors suggest Brown Bluff was initially erupted subglacially and later within an englacial lake. Field evidence suggests that there were drainage and refilling episodes of the lake, with drawdown and rise of at least 100m, during construction of the volcano (Fig. 3.9). This is characteristic of englacial lakes, which are typically unstable, especially in active volcanic regions [*Bjornsson*, 1988]. Changes in sea level or a non-glacial lake level of this magnitude are unlikely to have occurred during construction of the volcano. There is also no likely palaeotopography in this area which could have caused a lake to pond.

The total thickness of >400m for the Brown Bluff succession implies that the ice thickness was at least of a similar order. Skilling [1994] suggested that the original diameter of the volcano was about 12-15 km, probably with a single central vent. The rock is exposed in 5 corries (cirques) in the northern half of the volcano, labelled corries 1 to 5 (Fig. 3.8). The southern half of the volcano is obscured by ice. Brown Bluff conforms to the simple four-fold structural and lithological subdivision of tuyas, described by Jones [1969, 1970] form Laugarvatn, Iceland. This sequence comprises basal pillow lavas overlain by lapilli-sized (2-64mm) vitric tuffs (hyalotuffs) and is capped by hyaloclastite delta deposits and overlying subaerial lava flows. The following is an account of the processes and products that gave rise to this sequence at Brown Bluff. A more detailed account of the sedimentology is given in Skilling [1994].

Brown Bluff is subdivided into four stages: pillow volcano, tuff cone, slope failure and hyaloclastite delta/subaerial, and into five structural units (A to E, Table 1). Unit A is the basal and oldest unit, termed the pillow volcano stage and is composed dominantly of pillow lavas draped by massive hyaloclastite. Units B and D comprise the tuff cone stage and consist dominantly of vitric lapilli tuffs. These units are separated in the north-east of Brown Bluff by deposits of Unit C. This unit represents the slope failure stage, and consists dominantly of debris avalanche deposits. Unit E is the uppermost hyaloclastite delta/subaerial stage, and is composed of hyaloclastite delta deposits and overlying subaerial lava flows. Figure 3.9 summarizes the evoltion of Brown Bluff, including periods of lake drainage and recharge.

The major lithofacies and their relative abundances within each unit are illustrated in table 1. More detailed information on lithofacies is given in Skilling [1994]. Most of the facies are subdivided into "hyaloclastite" and "hyalotuff". These are defined on the basis of an estimate of the vesicularity of the lapilli (or coarser) clasts (<25%: hyaloclastite; >25%: hyalotuff). No particular fragmentation is implied by these terms. The sedimentary term "conglomerate" is used were

lithification prior to resedimentation had taken place. Sedimentary grain sizes are used throughout.

3.5.1. Interpretation of Unit A

Unit A has similarities to pillow volcano or seamount sequences, which often consist of a core of pillow lavas draped by massive and steeply bedded hyaloclastites [*Lonsdale and Batiza*, 1980; *Staudigel and Schmincke*, 1984]. The presence of pillow lavas at the base of the Brown Bluff succession and at the base of several other subglacial tuyas [Jones, 1969, 1970; Wörner and Viereck, 1987] suggests suppressed explosivity due to the overlying water/ice hydrostatic head. Unit A was constructed both by extrusive and intrusive processes with intrusion of pillows and pillow-margined bodies continuing at depth while tuff cone construction had already commenced [*Skilling*, 1994]. The massive hyaloclastites which drape the pillow lavas probably represent slumped deposits produced by failure of a steep-sided pillow volcano, or may represent slumped hyaloclastite delta deposits, from an early stage of emergence, developed without an intervening tuff cone.

3.5.2. Interpretation of Units B and D

Units B and D represent the deposits of a Surtseyan-type tuff cone. In the summit region, the presence of some inward-dipping beds, more pervasive hydrothermal alteration and radial dips of the outer cone slope deposits away from this area, imply the proximity of a vent. A steep inward-dipping contact in this region is interprted as a crater margin (Fig. 7 in *Skilling*, 1994), and the facies as the crater-fill. Tuff cone deposits were probably generated initially by subaqueous and subaerial wet tephra fallout (and surge?), with redeposition into the crater by mass flows and slumps. The occurrence of massive tephra in this area is interpreted as a vent slurry (Fig. 3.10). Curvicolumnar-jointed lavas in the summit area are interpreted as water-cooled lavas ponded within the crater. The interbedding of air and water-cooled lavas in the uppermost parts of unit D implies periodic water flooding of the crater floor.

The deposits of the outer cone slope were interpreted by Skilling [1994] as density-modified grain flows which had transformed to turbidites on the lower slopes. The occurrence of turbidite deposits, the absence of fluidal gravity flow deposits and the lack of reworking of the tephra, imply that the tuff cone was constructed within a ponded water environment (Fig. 3.10). The lateral impersistence and amalgamation of most of the beds suggests that the primary emplacement was from low-volume, discrete, simultaneous eruptions, interpreted as Surtseyan "tephra jets" (Fig. 3.10). There is no evidence of direct deposition by pyroclastic surges or airfall, implying either that all the tephra was resedimented downslope or back into the crater, or that the subaerial upper part of the tuff cone has been lost to erosion.

3.5.3. Interpretation of Unit C

The conglomerates of this unit record the failure of the north-east flank of the volcano, during tuff cone construction. Jigsaw-brecciated clasts, very poor sorting and poor mixing of clast populations with steep internal contacts, imply that this unit comprises dry debris avalanche deposits. This means that the ponded water necessary for turbidite deposition in Unit B had drained away, but that the lake must have refilled to allow continued depostion of Unit D turbidites (Fig. 3.9). The origin of the collapse is unclear, but the abundance of strongly hydrothermally altered clasts suggests that pervasive alteration of tephra to clay minerals in the vent region was a contributing factor. Drainage of the lake may also have contributed to collapse.

3.5.4. Interpretation of Unit E

The wedge-shaped morphology, steep to asymptotic bedding and the capping of lava flows with lobes which extend down into hyaloclastite [*Skilling*, 1994], suggest that Unit E represents hyaloclastite delta deposits (Fig. 3.11). The angle of deposition and the presence of inverse grading implies that the dominant process of deposition was density-modified grain flow. The occurrence of several conglomerate facies on the upper parts of the delta slope, and their juxtaposition with facies at the delta brink point from which they were derived, suggests that they originated by oversteepening and collapse in this area (Fig. 3.11). Skilling [1994] suggested that a variable input rate and volume of lava streams into the lake influenced the type of deposits that ponded at the brink point, and consequently the type of flows which are generated by their collapse. Narrow lava streams generated density-modified grain flows of cobble hyaloclastite, whilst wider streams formed pillow lavas which ponded at the brink point, and subsequently collapsed down the delta front. The widest streams of lava ponded at the brink point as large masses of curvicolumnar-jointed lavas, which also collapsed downslope. Rockfalls from the surrounding lithified tuff cone deposits also collapsed onto the delta. The juxtaposition of subaerial lava flows and hyalotuffs in corrie 4 without any intervening hyaloclastite (Fig. 7 in Smellie and Skilling, 1994), implies that a second drainage of the lake took place, allowing subaerial lava flows to fill a drained trough around the tuff cone (Fig. 3.9).

3.6. LAHARS AND JÖKULHLAUPS

As we've learned in the previous sections, hyaloclastic ridges and tuyas are vent phenomenon and are restricted to eruptions beneath ice sheets or water (Surtsey is a marine tuya); therefore they are a definitive

indicator of volcano and ice/water interaction. Floods and resulting mass flows are triggered by numerous causes including volcano-ice interaction, which form lahars and jökulhlaups. Although mass flows are not definitive indicators of volcano-ice interaction, but they are the most widespread, voluminous, and destructive products of subice eruptions.

Mass flows are gravity-driven, viscous non-Newtonian fluids. Features shared by most mass-flow deposits are poor sorting [*Rodine and Johnson*, 1976], lack of internal stratification [*Crandell and Waldron*, 1956], support of large clasts in a finer-grained matrix [*Johnson*, 1970], sharp contacts [*Hooke*, 1967], and form somewhat uniform thickness over large expansive flats [*Bull*, 1972] like the Martian lowlands. Morphologically, they may have steeply dipping lobate snouts, elevated margins, and in some cases central plugs of coarse clasts [*Johnson*, 1970]. Lahars have been known to abraid underlying bedrock, contain well-rounded cobbles and boulders with increased distance downstream, and deposit boulder clasts at flow obstructions [*Pierson et al.*, 1990]. The following discussion provides details of terrestrial lahars and jökulhlaups.

3.6.1. Alpine situations

3.6.1.1 Snowmelt by pyroclasts.

Explosive eruptions and subglacial melting at snowclad volcanoes swiftly generate in several ways [*Major and Newhall*, 1989] huge floods that are hazardous and can alter landscape. A common process is turbulent hot flows entraining snow to produce diverse "mixed avalanches" or debris-bearing snow avalanches, snowflows, slushflows, that may transform downvalley into great debris-laden floods [*Pierson and Janda*, 1994; *Pierson*, 1995].

During the first two minutes of the 18 May 1980 eruption of Mount St. Helens, a great turbulent pyroclastic surge melted snowpack, firn, and the surfaces of small glaciers. Thousands of small slushflows mixed with surge debris and accumulated unstably on steep slopes, coalescing downslope into slushy floods. Because constituent snow and debris retarded turbulence, the flows accelerated to 100 km/hr and more [Waitt, 1989] into valleys, which channeled the floods for tens of kilometers [*Pierson*, 1985; *Scott*, 1988]. During eruptions at Mount St. Helens between 1982 and 1991, small pyroclastic flows and surges repeatedly melted snowpack and formed various mixed pyroclast-snowgrain flows and floods (Fig. 3.12; [*Waitt et al.*, 1983; *Waitt and MacLeod*, 1987; *Pierson and Waitt*, 1997].

Alaska's snowy volcanoes often generate floods during eruption. Pyroclastic-flow deposits of eruptions of Augustine volcano in 1976 and 1986 grade into bouldery flood deposits [*Waitt and Begét*, 1996]. During each of three eruptions of Crater Peak (Mount Spurr) volcano in 1992, hot pyroclasts melted snowpack to form diverse debris-rich snow avalanches and flows intermediate between pyroclastic flows and lahars [*Waitt*, 1995].

3.6.1.2 Influence of glacier.

Like Nevado del Ruiz (section 3.3), other historical summit eruptions have triggered remarkable lahars. In winter 1989-90, dome-collapse pyroclastic flows at Mt. Redoubt, Alaska, stripped snowpack and firn to form a 3- to 20-m-thick icy diamict of snow grains and coarse ash freighted with huge blocks of andesite, glacier ice, and snow [*Waitt et al.*, 1994]. Some icy flows reached 14 km laterally over an altitude drop of 2.3 km. Erupting hot andesite probably mixed turbulently with snow, triggering avalanches that rapidly entrained firn and ice blocks from the heavily crevassed glaciers. Successive eruptions melted snow and glacier ice, incising and eventually beheading Drift glacier [*Trabant et al.*, 1994]. Resultant floods swept down Drift River valley for 35 km to the sea with discharges as high as 10,000 to 25,000 m^3/s and perhaps higher [*Dorava and Meyer*, 1994].

Glaciated Mount Rainier volcano has shed at least 60 sizable lahars in postglacial time, many of them clearly originating by eruption but a few very large ones originating as landslides, perhaps noneruptively [*Crandell*, 1971; *Scott and others*, 1995]. Though Rainier's eruptions tend to be far less explosive than Mount St. Helens, meltwater from pyroclastic eruption or geothermal heating near Rainier's summit must flow at least 5 km across or beneath steep, snowclad, much-crevassed glacier(s), a situation ripe for generating lahars and floods.

3.6.2. Icecap-icesheet situations

Many volcanoes, especially in middle to high latitudes, are crowned by icecaps or icesheets. Besides the swift melting of snow and ice during explosive eruption, these volcanoes and their geothermal systems can slowly melt ice and store water at the glacier bed ([*Björnsson*, 1975]; Fig. 3B in *Björnsson*, 1988). Such unstably icebound water may then break out as a subglacial jökulhlaup.

Iceland's Vatnajökull (icecap) overlies several hydraulic basins [*Björnsson*, 1988, plate 8], some of which are repeatedly swept by jökulhlaups. Archtypical jökulhlaups issue from the iceshelved caldera of Grímsvötn and flow down Skeidarársandur (fig. 3.1.A), typically without eruption. Geothermal heating gradually melts ice and lifts the lake level in Grímsvötn until hydraulic stability is exceeded and the lake suddenly drains subglacially [*Björnsson*, 1974; *Nye*, 1976]. Before 1940 the lake drained every 7-10 years and released large volumes (4.5 km^3) of water in large-discharge (>25,000 m^3/s) floods; in 1938 a flood was thought to

have peaked at about 30,000 m^3/s [*Björnsson*, 1992; *Gudmundsson and others*, 1995] or even 40,000 m^3/s (*Arni Snorrason*, pers. commun., 1998). Since 1940 jökulhlaups recur every 4-5 years, have typically smaller volume (1 to 2.5 km^3) and have lower peak discharges of 600 to 11,000 m^3/s. The unusually large 1996 Skeidará jökulhlaup of 52,000 m^3/s [*Jónsson et al.*, 1998a,b] was caused not just by passive geothermal melting but by a subglacial Gjálp eruption (section 3.4) that rapidly melted glacier ice.

Skafá River on southwest Vatnajökull has flooded suddenly many times. Some 40 km above the glacier terminus where the flood emerges, the glacier surface sags as two deep cauldrons, where Björnsson [1997] concludes geothermal heating melts the glacier bed to form cupola lakes. Sudden draining of either of these unstable lakes causes the ice surface directly above it to subside. Glacier flow then gradually infills the cauldrons.

In southeast Vatnajökull, icecapped Öræfajökull--Iceland's largest volcano--erupted explosively in 1362, sending huge debris-laden floods down outlet glaciers and erasing farmsteads [*Thorarinsson*, 1958]. Bouldery diamict hummocks 2-4 m high stand well out in front of the glacial moraines. Voluminous snow and glacier-surface ice may have been melted swiftly as at Mount St. Helens, but the floods were so large as to indicate also release of subglacially stored water as in 1918 at Katla.

Jökulsá á Fjöllum, a large glacial river, drains from Vatnajökull icecap 190 km northward to the sea. A scabland-carving flood had swept down lower Jökulsá á Fjöllum about 8000 years ago. About 2000 years ago Kverkfjöll caldera generated a gigantic flood that carved recessional cataracts and other scabland topography and laid gravel bars along 150 km of Jökulsá valley (Fig. 3.12). Smaller floods from Kverkfjöll in the 15th to 18th centuries destroyed farms in Axarfjördur. Many erosional and depositional features along this path resemble parts of Washington's Channeled Scabland, carved by Earth's largest (Pleistocene) floods. The late Holocene Jökulsá flood probably peaked between 0.4 and 1 million m^3/s [*Tómasson*, 1973; *Waitt*, 1998]. Subglacial Kverkfjöll caldera is a flood source, but farther west a subglacial fissure system and Bárdarbunga caldera are also potential sources. Had the 1996 Gjalp eruption (section 3.4) been farther north along the subglacial rift, its meltwater would have drained north to Jökulsá á Fjöllum rather than as it did south to Grímsvötn and Skeidará.

In south Iceland Myrdalsjökull (icecap) overlies Katla caldera, whose eruptions triggered enormous floods in 1660, 1725, 1755, and 1918 [*Jónsson*, 1982]. For the great 1918 Katla flood, Tómasson [1996] recalculates peak discharge at about 300,000 m^3/s--one of Earth's largest historic floods. It seems impossible that the enormous Katla jökulhlaups could have originated solely from melted snow and ice during brief eruption. Such volume so quickly implies breakout of dammed water, probably geothermally melted water stored in a cupola at the base of the ice. The 18th-century jökulhlaups of Mýrsdalssandur transported megaclasts, which Jónsson [1982] interprets to be deposits of dense debris flows. The poor sorting and weak stratification indeed resemble deposits of debris flows and hyperconcentrated flows from Mount St. Helens during its 18 May 1980 eruption [*Scott*, 1988].

3.7. PHREATIC CRATERS ON EARTH AND MARS

Another type of volcano/ice interaction occurs not as a true volcanic eruption, but as a result of steam explosions when lava flows over ground saturated with water or ice. Many of the Earth's volcanic regions are dotted with craters formed by steam explosions. Eruptions near the ocean or lava flowing into the sea commonly produce tuff rings or littoral cones. The Hawaiian and Galapagos islands contain many such features. The encounter of magma with ground water at depth can yield spectacular maars and explosion-collapse depressions such as the Pinacate craters of Sonora, Mexico.

Tuff rings and maars are found in Iceland, but a third type of steam explosion crater seems unique to that country. This is the phreatic crater, or pseudocrater, a landform resulting from explosions caused by lava flowing over water-saturated ground [*Thorarinsson*, 1960]. Phreatic craters are cones, generally resembling small cinder cones, which are rootless. That is, they are not formed above a vent.

The type locality for these pseudocraters is Lake Myvatn in northern Iceland, where basaltic lava flowed into the shallow lake basin around 2500 years ago. The southern shore and islands of Myvatn are dotted with over 1000 of these small cones (Figure 8, p. 204, in *Allen* 1979b). They can be as high as 30 m and have basal diameters as large as 320 m. The average Myvatn phreatic crater rises no more than 5 to 10 m and is approximately 50 to 100 m across. They were formed in a pahoehoe flow that ranged from 3 to 5 meters thick.

Though the phreatic craters at Myvatn are the best known, similar features occur in many parts of Iceland. At least eight fields of these craters have been mapped in other parts of the country. Phreatic craters are diverse in terms of composition, size and morphology. A broad central pit is common, often approximately one-half of the basal diameter across. Not even the crater form is requisite, however. Some of the features in southern Iceland have no summit pits, but are tumuli several

meters in height. Most phreatic craters are composed primarily of spatter, indicating that the explosions occurred while the bulk of the lava was still molten or at least semi-plastic. In some cases coherent basalt layers over a meter thick were uplifted and contorted without breaking [*Allen*, 1979b].

Individual phreatic craters are similar in many respects to small cinder and spatter cones, though their origin is significantly different. Morphologic evidence for an origin by steam explosion includes small size, large crater diameter/basal diameter ratio, lack of alignment among craters and confinement to a single flow.

The formation of phreatic craters by steam explosion is well accepted, on the basis of field observations. Fluid basalt can flow rapidly enough to trap water, which flashes to superheated steam. The calculated steam pressure is around 100 times the pressure exerted by the overlying lava [*Allen*, 1979b]. Violent explosions should be expected in such an environment.

What if, instead of water, lava flows over ice? The situations are actually very similar. The heat needed to melt ice and create superheated steam is only around 7% more than the energy required to make the steam by heating water [*Allen*, 1980]. If the ice is buried under a few meters of crushed rock, the heat of a lava flow is still sufficient to produce a steam explosion [*Allen*, 1979b]. This simplified description of the near surface may apply to some areas of Mars. With surface gravity less than 40% of that on Earth and a much more tenuous atmosphere, phreatic craters several times larger than their terrestrial counterparts might be formed [*McGetchin et al.*, 1974].

Groups of small cones, suggested to be of phreatic origin, have been recognized by several independent investigators in the Chryse Planitia [*Greeley and Theilig*, 1978], Deuteronilus Mensae [*Lucchitta*, 1978] and Acidalia Planitia [*Frey et al.*, 1979; *Allen*, 1980] regions of Mars (the origin of these features are under debate). Some of the Acidalia cones are shown in Figure 3.14. Most of these features measure approximately 600 m across the base, or double the size of the largest Myvatn crater. The martian examples are simple, occasionally overlapping cones with craters about half their basal width. The cones are morphologically distinct from primary and secondary impact craters of the same size, as well as from the small shield volcanoes in the region. If these are really phreatic craters they hold promise for mapping near-surface martian ice deposits.

3.8. HYALOCLASTIC RIDGES, TUYAS, LAHARS, AND JÖKULHLAUPS ON MARS

Possible analogs to terrestrial hyaloclastic ridges, tuyas, lahars, and jökulhlaups have been identified on Mars in the Acidalia Planitia region, north and northwest of Elysium Mons, a large volcanic construct on the east edge of the Utopia Planitia basin, and in Ares Valles at the site of the Mars Pathfinder Lander.

3.8.1 Hyaloclastic ridges and tuyas

Hodges and Moore [1978] noted that landforms resembling tuyas are concentrated in Acidalia Planitia. Allen [1979a] identified three tuya analogs and two possible hyaloclastite ridges in Acidalia. These features closely resemble Icelandic subice volcanoes in both morphology and size. Figure 3.15 shows a steep-sided, flat-topped mountain near latitude 45° N and longitude 21° in central Acidalia. A large cone with a shallow crater caps the mountain. The main plateau measures approximately 7 km by 4.8 km. By analogy, tuyas with similar structural features in Iceland are characteristically less than 10 km across. On Earth the height of the plateau approximates the local thickness of the glacier during eruption, generally 200 to 1000 m. Figure 3.14 features a sharp ridge at latitude 38° N and longitude 13°. The ridge is 6 km long and 2 km at maximum width. Several smaller ridges and fissures parallel the main feature. These dimensions are well within the range of the myriad morphologically similar hyaloclastite ridges throughout central Iceland. These subice eruptives range from <1 km to over 35 km in length and are generally 200-400 m high.

Northwest of Elysium Mons, landforms interpreted to be subice volcanoes occur as ridges on the flanks of the volcano [*Anderson*, 1992; *Chapman*, 1994] and as mesas in Utopia Planitia, near the Viking 2 Lander Site [*Hodges and Moore*, 1978; *Allen*, 1979a]. The enigmatic ridges and mesas on the flank of Elysium Mons were initially included in large areas mapped as lahars [*Christiansen and Greeley*, 1989; *Mouginis-Mark*, 1985]. However, as discussed in section 3.7, mass flows tend to form deposits of uniform thickness, not high-standing mounds and ridges. A subice volcanic origin of the ridges on the Elysium flank is supported when these features are compared with the Icelandic subglacial volcano Herdubreidartögl, a tuya whose narrow central ridge is clear evidence of its origin by fissure eruptions (Fig. 3.16.A; *Werner et al.*, [1996]). The rough-textured martian mounds and ridges contain similar central narrow linear ridges and pits; narrow ridges also extend away from the mounds (Fig. 3.16.B). Moberg ridges erupted along fissures southwest of the Vatnajökull ice cap (Fig. 3.17.A) form lineaments that appear very similar to some rough-textured ridges in the Elysium area (Fig. 3.17.B). Central,

narrow linear ridges and pits within the mounds and the alignment of mounds with Elysium Fossae (volcano-tectonic depressions) and linear chains of domes attest to fissure eruptions as the likely source of the mounds and ridges. Their rough texture, topography, and similarity to Icelandic features suggest that they may be formed from friable materials such as palagonite and breccia and are possible hyaloclastic ridges. The slopes of the Elysium ridges average about 2°. Curiously, the slope of Kalfstindar (Iceland, fig. 3.5) is a much steeper 18°. This discrepancy may be due to erosion of friable hyaloclasts on the much older martian ridges (<0.8 ma for Kalfstindar vs. 200 ga or more on Mars) and because Kalfstindar is unusually large (sec. 3.5).

Mesas in the northeastern part of Utopia Planitia may be tuyas erupted from central point sources (Fig. 3.18.A; [*Hodges and Moore*, 1978; *Allen*, 1979a]). Gaesafjöll (Fig. 3.18.B) is an Icelandic tuya that erupted palagonite-forming basalt from a central source until the surrounding ice cap retreated and the meltwater drained; continued eruptions produced layered, subaerially extruded lavas and a central summit crater [*Van Bemmelen and Rutten*, 1955]. One mesa at about lat 46° N., long 229° (Fig. 3.18.A) is superposed on polygonal terrain; the break in slope between the cap rock (layered lavas?) and underlying slope (palagonite?) is clearly visible. The morphology of the mesas, their location in the Utopia Basin near the large volcanic center of Elysium Mons, and their alignment with other volcanic cones [*Chapman*, 1997] suggest that they may be tuyas.

As noted in the introduction, water may have ponded in the northern lowlands of Mars, which includes Utopia Planitia. As some additional evidence suggests, any hypothetical paleolake or ocean (and any connecting ephemeral lakes) in the basin was likely frozen to some depth [*Lucchitta et al.*, 1986, 1987; *Scott and Underwood*, 1991; *Kargel and Strom*, 1992]. The thickness of the ice sheet and water level may be estimated from the height of the subglacial volcanoes (section 3.5) in Utopia Planitia. Photoclinometric measurements of the ridges and mesas northwest of Elysium indicate that the paleo-ice sheet in Utopia may have been 200 meters thick [*Chapman*, 1994].

3.8.2 Lahars, and jökulhlaups

Surrounding and downslope of the hypothetical hyaloclastic ridges of northwest Elysium Mons, extending 1,500 km away into central Utopia Planitia, is a lobe of coarse, rough-appearing material with inner, smooth areas. This lobe was interpreted by Christiansen [1989] to be laharic material, based on its morphologic similarity to terrestrial mass-flows. Because of its proximity to the hypothetical hyaloclastic ridges, this rough-textured material was reinterpreted as jökulhlaup deposits

[*Chapman*, 1994]. The material appears to have been discharged from fissures related to the Elysium moberg ridges, as the lobe deposit can be traced back to these volcanic features. These rough deposits form topographically lower, smoother appearing deposits locally bounded by lobate scarps. The smooth areas contain nested 1-km-wide concentric craters and smooth-floored, irregular channels that may be due to collapse and dewatering of mass flow material [*Christiansen*, 1989].

On the north flank of Elysium Mons (west of Hecates Tholus) are local, subdued relief flows that are less lobate than other upflank lava flows. Mouginis-Mark [1985] believes the subdued flows might be the martian equivalent of Icelandic jökulhlaups. Finally, deposits in Ares and Tiu-Simud Valles at the site of the Mars Pathfinder Lander were compared by Rice and Edgett [1997] to similar Icelandic outwash plains formed by jökulhlaups.

3.9. VOLCANO-GROUND WATER INTERACTIONS AND SNOW/ICE PERTURBATIONS ON MARS

Ground ice on Mars is a likely indicator of ground water at depth. In fact, volcano-ground water or ground ice interactions appear to have been pervasive throughout the geologic history of Mars. Ground water or ground ice can interact with volcanoes in several ways. Shallow intrusions can interact directly with ground water, resulting in explosive volcanism, such as maar craters. Or water might enter a magma chamber, thereby affecting the style of volcanism, perhaps producing pyroclastic activity (See Chapters 2 and 4). A magmatic intrusion at depth will alter the flow of ground water, producing convecting cells of water and a potentially long-lived hydrothermal system.

Perhaps the most convincing evidence for volcano-ground water interactions are the large channels found immediately adjacent to some volcanoes. These channels emanate from discrete collapse zones, suggesting a close relation between intrusive igneous activity and the sudden release of ground water. Examples are the channels west of Elysium Mons volcano [*Mouginis-Mark*, 1985; labeled 10 in fig. 4.1], Dao Vallis on the flank of the volcano Hadriaca Patera [*Squyres et al.*, 1987; fig. 4.13], and the small channels immediately to the east of Olympus Mons [*Mouginis-Mark*, 1990; fig. 4.2]. On the largest possible scale, Baker et al. [1991] have suggested that the formation of the largest outflow channels was linked to a regional-scale hydrothermal system triggered by the formation of the Tharsis bulge.

On a smaller scale, fluvial valleys are found predominantly in the ancient cratered highland terrain and upon the slopes of some volcanoes. The northern flank of the very young volcano Alba Patera is dissected by

many small valleys [*Gulick and Baker*, 1989; 1990]. Several other volcanoes, including Apollinaris Patera, Hecates Tholus, and Ceraunius Tholus also exhibit small valleys. These valley systems conceivably could have been formed by rainfall processes, however evidence exists for ground water outflow in many cases. Furthermore the volcano Alba Patera formed long after the putative early warm, wet period of Mars' climatic history. Fluids discharged by hydrothermal activity are thus an attractive alternative for the source of these valleys [*Gulick and Baker*, 1989; 1990].

3.9.1. Hydrothermal Systems

The possible influence of hydrothermal systems has long been recognized. Schultz et al. [1982] and Brakenridge et al. [1985] suggested that impact-induced hydrothermal systems could be responsible for valleys on ejecta blankets on Mars. Gulick and Baker [1989, 1990] proposed that the discharge of hydrothermal fluids to the surface was an important process for the formation of those valleys which formed on the flanks of martian volcanoes, particularly on the younger, ash covered Hesperian and Amazonian aged volcanoes. Some hybrid models include both precipitation and hydrothermal systems. Gulick et al. [1997] proposed that a substantial amount of water could be transported during modest greenhouse periods from surfaces of frozen bodies of water to higher elevations, despite global temperatures well below freezing. This water, precipitated as snow, could ultimately form fluvial valleys if deposition sites were at or near regions of hydrothermal activity. Hydrothermal systems have also received a great deal of attention as agents for producing aqueous alteration of the shergottite-nakhlite-chassignite (SNC) meteorites [*Wentworth and Gooding*, 1994], for exchanging deuterium and hydrogen between the crust and atmosphere [Jakosky and Jones, 1994], for providing paleohabitats for life, and for preservation of fossils [*Walter and DesMarais*, 1993; *Farmer and DesMarais*, 1994].

An example of an idealized model of ground water flow associated with a magmatically generated hydrothermal system as might be produced by the formation of a martian volcano is shown in Figure 3.19. Soon after emplacement of the magma, the outer shell of the magma chamber starts to solidify, forming a low permeability outer shell of hot rock, the thickness of which increases with time [*Gasparini and Mantovani*, 1984]. Thermal energy is transported primarily by conduction from the magma through the shell, and then primarily by convection into the saturated, permeable country rock. This shell prevents ground water from contacting the intrusion itself. Surrounding ground water that is heated by the magma forms an upwardly moving buoyant plume near the intrusion. Colder, denser ground water flows in toward the intrusion from surrounding regions and continues to replace the upwardly moving ground water as long as a thermal gradient exists. Depending on the size of the intrusion, ground water within several tens of kilometers or more could flow into the system. Ground ice above and near the intrusion would be melted, locally eliminating or thinning the permafrost zone [*Gulick and Baker*, 1992; *Gulick*, 1993]. Ground water reaches the surface as liquid or vapor or both. The near-surface behavior of the hydrothermal fluids would depend on their temperature and mineral concentration, and on the atmospheric temperature and pressure.

Ground water that reaches the near-surface environment can contribute to the geomorphic modification of that surface. If local hydrologic and lithologic conditions permit, water could flow on the surface and re-enter the ground water system in regions where the rate of infiltration is sufficiently high. However, if the atmospheric temperature and pressure are not favorable for fluid flow, ground water would initially start to boil and evaporate but then freeze due to the heat-liberating process of evaporation. This process would result in the formation of an insulating ice layer beneath which subsequent outflows of hydrothermal water may move as ice-covered streams [*Wallace and Sagan*, 1979; *Carr*, 1983; *Brakenridge et al.*, 1985]. Ground water that does not interact with the surface would help to recharge near-surface aquifers and eventually could outflow to the surface farther away from the intrusion. Whether ground water remains liquid would depend on the local lithologic conditions, the temperature and mineral concentration of the water, and the atmospheric conditions at the time of ground water outflow.

3.9.2. Permafrost

The presence of ice-rich permafrost would affect an active hydrothermal system on Mars. Upwardly moving hydrothermal fluids must melt through permafrost to reach the surface. Assuming that convective warming is much faster than conductive cooling, Gulick [1998] computed the time required to melt through a 2-km-thick permafrost zone. She found that if permafrost on Mars fills a region with a pore space of 10 to 35%, then hydrothermal fluids can melt through the permafrost in several thousand years. This time is short compared to the lifetime of moderately sized hydrothermal systems (about 100,000 years for a 50 km^3 intrusion). Therefore, the presence of an ice-rich permafrost zone should have a negligible effect on the lifetime of a hydrothermal system on Mars.

Even directly above a volcanic intrusion, the surface temperature at Mars would be primarily controlled by the balance between absorbed

solar and emitted infrared radiation [*Fanale et al.*, 1992]. The surface temperature will remain below freezing and a residual ice-rich permafrost zone will remain near the surface, except for areas directly above the intrusion or where springs or seeps have formed. Figure 3.20 (from Gulick, 1998) illustrates the equilibrium thickness for a variety of heat flows and surface temperatures (applying McKay et al. [1985], equation 2). The present-day background geothermal gradient probably provides on average around 0.03-0.04 W/m^2 [*Fanale*, 1976; *Toksîz and Hsui*, 1978; *Davies and Arvidson*, 1981]. However, the average heat flow in the presence of an active terrestrial hydrothermal system can range from 2 to 5 W/m^2. For example, at Wairaki, New Zealand, the regional heat flow averaged over approximately 50 km^2 is 2.1 W/m^2, whereas fluxes averaged over the more intense regions are of order 500 W/m^2 particularly in localized areas around springs [*Elder*, 1981]. The graph shows that the equilibrium permafrost thickness above an active hydrothermal system may be less than 100 m. Any inhomogeneities in the subsurface, such as fractures, would permit egress of hydrothermal water to the surface. The system can adjust to the new permafrost equilibrium thickness in less than about 10,000 years, a time that is short compared to the lifetime of the hydrothermal system (Fig. 3.20).

Though these theoretical ideas apply to hydrothermal systems on Mars, to date little work has been done to analyze specific potential hydrothermal sites, partly because relatively little imaging data at sufficiently high resolution is available to unequivocally identify sites of hydrothermal outflow. Also, insufficient high resolution spectral data exists to identify hydrothermal minerals that might be associated with sites of hydrothermal activity. Until such data become available, it will be difficult to uniquely constrain the properties of specific hydrothermal systems on Mars.

3.9.3. The aureole deposits of Olympus Mons

The origin of the lobate overlapping aureole deposits around Olympus Mons remains controversial. These deposits extend from the volcano's basal scarp to distances up to 1000 km [*Mouginis-Mark et al.*, 1992] and consist of enormous lobes of closely spaced, roughly parallel arcuate ridges. Data from the laser altimeter (MOLA) aboard Mars Global Surveyor indicate that the volcano's basal scarp appears to be approximately 8 km high and the aureole deposits are roughly 3 to 4 km in relief [*Smith et al.*, 1998]. Several hypotheses have been put forth to explain these landforms, including Hodges and Moore [1979], who suggested that Olympus Mons is analogous to Icelandic tuyas and its aureole deposits were formed by subglacial lava flows. They conclude that the associated ice sheet thickness was comparable to that of the aureole deposits, or 3-4 km thick. Difficulties with this hypothesis cast doubt on its validity, and mainly revolve around the preferential location of an extremely thick ice sheet late in martian history around Olympus Mons. More recently, Lopes et al. [1980] and Frances and Wadge [1983] have attributed the presence of the aureole deposits as mass movement of slide material from the volcano.

3.10. CONCLUSION

Volcano/ice interactions produce meltwater. Meltwater can enter the groundwater cycle and under the influence of hydrothermal systems, it can be later discharged to form channels and valleys or cycled upward to melt permafrost. Water or ice-saturated ground can erupt into phreatic craters when covered by lava. Violent mixing of meltwater and volcanic material and rapid release can generate lahars or jökulhlaups, that have the ability to freight coarse material, great distances downslope from the vent. Eruption into meltwater generate unique appearing edifices, that are definitive indicators of volcano/ice interaction. These features are hyaloclastic ridges or mounds and if capped by lava, tuyas.

On Earth, volcano/ice interactions are limited to alpine regions and ice-capped polar and temperate regions. On Mars, where precipitation may be an ancient phenomenon, these interactions may be limited to areas of ground ice accumulation or the northern lowlands where water may have ponded fairly late in martian history. The recognition of features caused by volcano/ice interactions could provide strong constraints for the history of volatiles on Mars

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Table Placement

Table Captions

Table 3.1 Description and interpretation of lithofacies of Brown Bluff Volcano, a tuya in Antarctica.

Figure Captions

Figure 3.1. (A) Distribution of Upper Pleistocene (<0.78 Ma) and Recent hyaloclastites and associated lithofacies in Iceland. Modified after Johannesson & Saemundsson [1998]. The hyaloclastite units shown cover an area of about 10.600 km2, those hidden by present ice caps may bring the total area to some 15.000 km2. The total volume is unknown, but by rough estimate, 60-70 % of the exposed units are primary or redeposited hyaloclastites, 15-25 % are pillow lavas, the reminder being subaerial lavas and intrusions. Basalts are ≥90 % of the volume, probably <5 % are intermediate rocks and ≤5 % silicic rocks. Insert shows the location of Gjálp and Grímsvötn. (B) Sections showing schematically the development of the subglacial ridge in the Gjalp eruption and the accumulation of meltwater in the subglacial lake of the Grímsvötn caldera (after *Gudmundsson et al.*, 1997). During the latter part of the eruption an ice canyon formed in the ice surface, where part of the meltwater flowed before disappearing down into the glacier. Five weeks after the eruption started, the subglacial lake was drained in a very swift jökulhlaup.

Figure 3.2. The eruption on October 3, 1996. The crater in the ice surface is about 500 m in diameter. The ice surface is covered with tephra (photo by M.T. Gudmundsson).

Figure 3.3. Schematic sections showing the penetration of the ice cap by the eruption. The central piston was rapidly melted as it collapsed onto the underlying vent [*Gudmundsson et al.*, 1997].

Figure 3.4. Morphology and main structural units of three Upper Pleistocene subglacial volcanoes in the region southwest of Langjökull, the Western Volcanic Zone in Iceland. A. Tindaskagi, and B. Skefilsfjöll, two typical medium-sized hyaloclastite ridges. Only minor pillow lava is exposed. The ridges are partly buried by Recent lava flows. C. Stóra-Björnsfell, an elongated medium-sized tuya. The basal pillow lava appears originally to have erupted from a fissure.

Figure 3.5. Schematic sections through the large Kalfstindar hyaloclastite ridge in the Western Volcanic Zone in Iceland. From Jones [1970].

Figure 3.6. An oblique aerial photo of subglacial hyaloclastite ridges southwest of Vatnajökull, Iceland. The ridges are 20-30 km long and have a relief of 300-500 m. Photo by Oddur Sigurdsson.

Figure 3.7. A diagram showing the relationship between what various authors have designated basaltic subglacial hyaloclastite ridges and tuyas, as regards length and maximum width. Data from the Western and Eastern Volcanic Zones in Iceland. The estimated position of the basalt-andesitic hyaloclastite ridge formed in the 1996 Vatnajökull eruption is indicated.

Figure 3.8. Maps illustrating location of Brown Bluff volcano. In B, the dips of hyalotuff cone deposits (Units B and D) radiate approximately from the present summit region, indicating a source region in that area

Figure 3.9. Summary diagrams of the evolution of Brown Bluff volcano. 1: entirely subglacial (pillow volcano stage); 2: hyalotuff cone stage, terminated by lake drainage (3); 4: lake refilling and continuation of tuff cone construction; 5: subaerial lava flows erupted and formation of hyaloclastite deltas; 6: second episode of lake drainage and partial infilling of trough surrounding the cone by subaerial lava flows.

Figure 3.10. Diagram illustrating the eruptive and depositional processes which operated during the hyalotuff cone stage.

Figure 3.11. Diagram illustrating the processes and products generated during hyaloclastite delta construction (Unit E).

Figure 3.12. Snowy flood from crater of Mount St. Helens induced by small explosive eruption in March 1982. From outer margin to center is the succession: snowflow, slushflow, watery flow.

Figure 3.13. Scabland with relief of 20 m carved about 2000 yrs ago by huge jökulhlaup along Jökulsá á Fjöllum, north Iceland.

Figure 3.14. Part of Viking Orbiter image 72A02 showing possible hyaloclastic ridge and surrounding phreatic cones in Acidalia Planitia, Mars; note central ridge and alignment with nearby fissures; illumination is from the left; scale bar equals 5 km.

Figure 3.15. Part of Viking Orbiter image 26A28 showing possible tuya in Acidalia Planitia, Mars; illumination is from the left; scale bar equals 5 km.

Figure 3.16. Comparison of Icelandic and Martian features; illumination is from the right; scale bars equal 5 km. (A) Herdubreidartögl [*Van Bemmelen and Rutten*, 1955], a tuya 4 km wide; its narrow central ridge is evidence of emission via fissure eruptions. (B) Part of Viking Orbiter image 541A10 showing a mound interpreted as hyaloclastic ridge; note mound contains an inner ridge.

Figure 3.17. Comparison of Icelandic and Martian features; illumination is from the bottom left; scale bars equal 5 km. (A) Hyaloclastic ridges, 2-3 km wide, erupted along fissures southwest of Vatna ice cap [*Van Bemmelen and Rutten*, 1955]. (B) Part of Viking Orbiter image

541A20 showing possible hyaloclastic ridges; note central pits and ridges extending away from mounds.

Figure 3.18. Comparison of Martian and Icelandic features; scale bars equal 5 km. (A) Possible tuya on Mars in Utopia Planitia at about lat. 46°N., long 229°; arrows denote break in slope between cap rock and underlying material; illumination is from the upper left. (B) Gaesafjöll [*Van Bemmelen and Rutten*, 1955], a tuya about 5 km wide with a central summit crater; illumination is from the top.

Figure 3.19. Conceptual model illustrating the groundwater flow field of a vigorous hydrothermal system associated with volcano formation on Mars. Vertical scale is exaggerated by a factor of 4 in order to illustrate details of groundwater flow in the stratigraphic layers of the volcano. Martian volcanoes are unusually large compared to their terrestrial counterparts, ranging on the order of 10^2 to 10^3 km in diameters, and those that are dissected by valleys tend to have low aspect ratios with average slopes less than a few degrees. Our numerical model considers hydrothermal systems associated with magmatic intrusions on Mars in general. Topography, multiple intrusions, and boiling are not directly simulated in our model, although their effects on these systems are discussed elsewhere in this paper. (Figure from Gulick [1993,1998]).

Figure 3.20. Equilibrium permafrost thickness as a function of geothermal heat flow. Permafrost thickness is shown for three mean surface temperatures. Typical estimates of current Martian heat flow as well as geomorphologic evidence indicate permafrost thicknesses of approximately 2 km. Higher surface temperatures and interior heat fluxes both produce thinner permafrost layers. Heat flow above an active hydrothermal system may be as large as 3 to 5 W m^{-2}, indicating equilibrium thicknesses of several hundred meters or less. (Figure from Gulick [1993,1998])