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Continental basaltic volcanoes — Processes and problems

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ABSTRACT

Monogenetic basaltic volcanoes are the most common volcanic landforms on the continents. They encompass a range of morphologies from small pyroclastic constructs to larger shields and reflect a wide range of eruptive processes. This paper reviews physical volcanological aspects of continental basaltic eruptions that are driven primarily by magmatic volatiles. Explosive eruption styles include Hawaiian and Strombolian (*sensu stricto*) and violent Strombolian end members, and a full spectrum of styles that are transitional between these end members. The end-member explosive styles generate characteristic facies within the resulting pyroclastic constructs (proximal) and beyond in tephra fall deposits (medial to distal). Explosive and effusive behavior can be simultaneous from the same conduit system and is a complex function of composition, ascent rate, degassing, and multiphase processes. Lavas are produced by direct effusion from central vents and fissures or from breakouts (boccas, located along cone slopes or at the base of a cone or rampart) that are controlled by varying combinations of cone structure, feeder dike processes, local effusion rate and topography. Clastogenic lavas are also produced by rapid accumulation of hot material from a pyroclastic column, or by more gradual welding and collapse of a pyroclastic edifice shortly after eruptions. Lava flows interact with — and counteract — cone building through the process of rafting. Eruption processes are closely coupled to shallow magma ascent dynamics, which in turn are variably controlled by pre-existing structures and interaction of the rising magmatic mixture with wall rocks. Locations and length scales of shallow intrusive features can be related to deeper length scales within the magma source zone in the mantle. Coupling between tectonic forces, magma mass flux, and heat flow range from weak (low magma flux basaltic fields) to sufficiently strong that some basaltic fields produce polygenetic composite volcanoes with more evolved compositions. Throughout the paper we identify key problems where additional research will help to advance our overall understanding of this important type of volcanism.

1. Introduction

Monogenetic basaltic volcanoes are the most common type of continental volcano. They range in form from small scoria cone volcanoes where much of the material was erupted by explosive mechanisms with variable proportions of lava flows (e.g., Wood, 1980; Gutmann, 1979; Luhr and Simkin, 1993; Gutmann, 2002; Valentine et al., 2005; Martin and Nemeth, 2006; Valentine et al., 2006, 2007), to small shields and chains of shields (e.g., Kuntz et al., 2002; Hughes et al., 2002a). Basaltic volcanoes are most common in extensional or convergent tectonic settings, or in intraplate settings that are near active regions (e.g., Colorado Plateau, U.S.A.), and occur as isolated features (e.g., Glazner et al., 1991), as basaltic volcanic fields (e.g., Condit et al., 1989; Connor, 1990; Camp et al., 1991; Briggs et al., 1994; Connor and Conway, 2000; Aranda-Gómez et al., 2003; Hare and Cas, 2005; Hare et al., 2005; Valentine and Perry, 2006), or in association with (often peripheral to) silicic volcanoes and calderas (e.g., Smith et al., 1970; Carn, 2000). Despite the information these volcanoes contain that is pertinent to our broader understanding of basaltic magmatism, the potential hazards they pose (e.g., Crowe, 1986; Siebe et al., 2004; Siebert and Carrasco-Núñez, 2002; Valentine, 2003; Houghton et al., 2006), and the fact that they are the most abundant subaerial volcanic feature, continental basaltic volcanoes have received relatively little attention in terms of eruption and emplacement processes compared to volcanoes with more evolved compositions.

In this paper, we review recent physical volcanological studies of monogenetic continental basaltic volcanoes and provide our perspective on key problems that need to be addressed by future research. Our focus is on "common" monogenetic basaltic volcanoes whose eruptions are driven primarily by magmatic volatiles. Many basaltic volcanoes are produced wholly or in part by explosive magma–water interaction (hydrovolcanism) to produce features such as tuff rings, maars, and tuff cones and their feeder diatremes. These have been described extensively elsewhere in the literature (e.g., Crowe and Fisher, 1973; Sheridan and Wohletz, 1981; Wohletz and Sheridan, 1983; Lorenz, 1986; White, 1991; Houghton and Schmincke, 1989;

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Houghton et al., 1999; Vazquez and Ort, 2006) and are therefore are not a focus in this paper.

We begin by discussing explosive eruptive processes of basaltic volcanoes and the resulting pyroclastic facies. This is followed by a review of basaltic lava emplacement processes and resulting features. Based upon observations at historic eruptions such as Parícutin (Krauskopf, 1948; Luhr and Simkin, 1993) and at prehistoric volcanoes such as Lathrop Wells (Valentine et al., 2007) it is clear that highly explosive eruptive activity can be accompanied by simultaneous lava effusion from the same subsurface plumbing system, so it is important to view the separation of pyroclastic and effusive processes in our discussion as one of convenience rather than as an inference that the two processes only occur separately in nature. Additionally, cone growth can be closely coupled to lava effusion through processes of rafting, and in turn, this process can depend upon the slope of the surface upon which a cone is built. Small shields (small in comparison to Hawaiian volcanoes, for example) are the effusive end member of the eruptive behavior discussed here, although even small shields commonly display some pyroclastic activity. Finally, we review field and theoretical aspects of the plumbing systems that feed the various types of basaltic eruption styles. Throughout, we point out issues that require additional research, with the hope that these suggested research topics will motivate work to greatly advance our understanding of continental basaltic systems.

2. Explosive eruptive processes and resulting pyroclastic facies

Focusing only on eruptions driven by magmatic volatiles, it is possible to describe four end-member behaviors that depend mainly on the mass flux and degree of fragmentation (or characteristic vesicle-size) of an erupting mixture, building on discussions such as found in Walker (1973a) and Parfitt (2004). These are: (1) Hawaiian, (2) Strombolian, (3) violent Strombolian, and (4) weak ash emissions. Strombolian and Hawaiian styles are tied to specific phenomena observed at the frequently erupting sites that are their namesakes. Sparks et al. (1997), Vergniolle and Mangan (2000), and Parfitt (2004) provide summaries of Hawaiian and Strombolian eruptive phenomena. Vergniolle and Mangan (2000) defined Hawaiian eruptions as relatively coarse-grained (<0.17% ash). Strombolian eruptions are characterized by discrete bursts that also are coarse-grained, but are different from Hawaiian eruptions mainly in mass flux (sustained fluxes of <50–1000 m^3/s for Hawaiian eruptions, whereas Strombolian activity ejects ~10^{-3}–100 m^3 per explosion with typically several explosions per hour). The third type of behavior — violent Strombolian — has a more confusing origin that is reviewed below. The fourth type, which we refer to simply as weak ash emissions, is relatively fine-grained with a low mass flux and negligible gas thrust as has been described in detail by Patrick et al. (2007), and is not further discussed here. The eruptive behaviors in their end-member form produce characteristic pyroclastic facies within and beyond the cones. A given volcano may be formed by any combination of these behaviors, as well as phases of hydrovolcanic activity (e.g., Houghton and Schmincke, 1989; Valentine and Groves, 1996), and therefore may contain different pyroclastic facies (Houghton et al., 1999; Valentine et al., 2005, 2006, 2007).

Over the decades, volcanologists developed a habit of using the terms Hawaiian and Strombolian to describe landforms. For example, scoria cones are commonly referred to as "Strombolian cones" (see, for example, Brenguier et al., 2006), and fissure-fed basaltic eruptions as "Hawaiian fissures." This mixed use of terminology has resulted in two unfortunate outcomes: (1) ambiguous terms such as "violent Strombolian," as discussed below; and (2) an assumption that a given volcanic landform, such as a scoria cone, is produced by a single, presumably well-understood, eruption process. This latter outcome is partly the cause for the relatively low level of attention paid by the physical volcanology community to continental basaltic volcanoes, compared to more silicic ones (see also Martin and Nemeth, 2006).

The term "violent Strombolian" is an example of the confusion caused by mixing terms for eruptive styles and landforms. MacDonald (1972) coined the term to describe cone-building eruptions that "project voluminous showers of scoria and bombs to heights of hundreds or thousands of feet, accompanied by a dense black ash cloud" and that "are sometimes of quite long duration." This eruptive style is not consistent with activity that is normally observed at Stromboli (e.g., Vergniolle and Mangan, 2000), but we infer that a prejudice toward associating scoria cones with Strombolian eruptions

Table 1
Facies characteristics of pyroclastic deposits from explosive, magmatic basaltic eruption styles

Eruptive style	Bedding	Texture and grading	Clast size	Clast shape	Clast vesicularity	Welding
Proximal (cone) deposits						
Strombolian	Mainly lenticular over several meters to ~10 m, decimeters to ~1 m thick	Reverse graded to massive	Coarse lapilli, blocks and bombs	Aerodynamic shapes (e.g., ribbon, spindle) of coarsest clasts, angular to slightly rounded smaller clasts	Moderate vesicularity with a wide range of sizes	Local moderate to dense welding very close to vent, non-welded elsewhere
Hawaiian	Lenticular to continuous over >10 s m	Massive to reverse graded	Coarse lapilli and bombs	Aerodynamic and fluidal (e.g., wrapping around underlying clasts), ragged margins	Moderate vesicularity with a wide range of sizes	Densely to partially welded over much of cone extent
Violent Strombolian	Mainly planar, continuous over >10 s m, locally lenticular. Localized, thin, ash-rich cross-bedded horizons	Massive to graded, internal planar stratification, local reverse- graded lenses reflecting grain avalanches	Coarse ash to coarse lapilli, sparse blocks and bombs	Mainly blocky and angular to slightly rounded. Sparse aerodynamic shapes	Moderately to highly vesicular with abundant small vesicles	Non-welded
Medial to distal (fallout) deposits						
Strombolian	Planar, localized within 100 s m of cone base	Massive to reverse or normally graded	Fine to coarse ash, sparse fine lapilli	Blocky to bubble wall shapes	Moderate to high	Non-welded
Hawaiian	Planar, localized within 100 s m to a few km from cone base	Massive to reverse or normally graded	Fine to coarse ash, lapilli close to cone	Small ribbons, aerodynamic shapes in lapilli, bubble wall shapes in ash, and reticulite	Moderate to extremely high (e.g., reticulite), large vesicles common in lapilli	Possibly partly welded close to cone
Violent Strombolian	Planar, extending km to a few 10 s km from vent. Localized ash-rich, cross-bedded horizons	Massive to reverse or normally graded	Fine to coarse ash and lapilli sizes	Small ribbon fragments and blocky to slightly rounded lapilli, bubble wall to blocky shapes in ash	Moderate to high with abundant small vesicles in lapilli (possible pumice texture)	Non-welded

partly led researchers toward maintaining a link to Stromboli in their terminology. Walker (1973a) continued this usage of the term violent Strombolian to describe eruption phenomena that produced widespread, ash-rich tephra fallout at Parícutin volcano (Mexico) and other eruptions that produced both scoria cones and significant fallout over distances of order 1–10 km and areas of 10–100 s of km^2.

Below, we summarize the characteristics of Strombolian, Hawaiian, and violent Strombolian eruptive behaviors and of the resulting pyroclastic deposits (proximal cone- forming deposits, and medial–distal fallout deposits; Table 1) in their end-member form. Strombolian and Hawaiian styles are very briefly summarized because overviews have been provided by many authors (e.g., Fisher and Schmincke, 1984; Cas and Wright, 1987; Francis 1993; Vergniolle and Mangan, 2000; Vespermann and Schmincke, 2000; Parfitt, 2004), compared to the violent Strombolian style which seems to lack an accepted definition; our purpose is mainly to set the stage for a comparative discussion of the three. Implicit in our discussion is that these three eruptive styles are driven primarily by magmatic volatiles rather than significant hydrovolcanic components. The reasoning behind this is discussed below in Section 2.4, which includes some thoughts on the relationships between the styles and issues that would benefit from further research. Grain size terminology follows the definitions of White and Houghton (2006).

2.1. Strombolian

Vergniolle and Mangan (2000) summarized the typical activity at Stromboli as characterized by low, temporally discrete fountains of incandescent clasts. The fountains are caused by the bursting of large gas pockets that have risen through a relatively stagnant or slowly rising column of magma in the feeder conduit. The bursts eject juvenile, fluid clots on ballistic trajectories at angles up to ~75° from vertical. McGetchin et al. (1974) described how the accumulation of these ballistic clasts builds a cone around the vent. Initially clasts are deposited onto the pre-existing surface but a raised rim rapidly develops at the distance of maximum clast accumulation. Repeated explosions continue to build this rim, and grain avalanching maintains the outer cone slopes close to the angle of repose while also recycling previously-erupted bombs (whole and broken into smaller lapilli and blocks) down the inner cone slopes back into the vent, to be re-erupted with some amount of juvenile material during subsequent bursts. As a result, proximal facies tend to be a mixture of roughly equant, angular to variably abraded lapilli and blocks along with fluidal or flattened bombs (Fig. 1A). Multiple episodes of avalanching and changes in explosion characteristics during a single eruption result in internal unconformities as the rim changes locations. Some bursts may eject sufficient juvenile material in low, short-lived fountains to result in locally high accumulation rate and welded to partly welded facies (e.g., Valentine et al., 2006) that are limited to tens of meters in lateral extent.

Generally, Strombolian explosions produce only minor convective ash clouds, with most of the clasts being in the lapilli to bomb size range (e.g., McGetchin et al., 1974, estimate ~50% of the material erupted at their study site was between 10 and 40 cm in size), and it is likely that much of the ash in such clouds is generated by abrasion of already-cooled and brittle clasts when they are recycled in the vent (see Patrick et al., 2007). Commonly, large blocks and bombs that land on the outer slopes roll downward to form a coarse-grained ring around the foot of a cone (e.g., McGetchin et al., 1974; Taddeucci et al., 2004). Because the ejecta clasts are relatively coarse, heat transfer to the erupting gas and potentially entrained air is limited; the combination of poor thermal and momentum coupling between clasts and gas act against development of a higher sustained eruption column. Most products of Strombolian eruptions are deposited in the proximal cone facies, with only limited fallout beyond the cones themselves (Table 1).

2.2. Hawaiian

Hawaiian-style eruptions have a large descriptive literature that we only touch upon here (see Vergniolle and Mangan, 2000; Parfitt, 2004, for more detailed reviews). As with Strombolian eruptions, clasts are relatively large and poorly coupled, both in momentum and thermal energy, with erupting gas. The critical difference between the two eruption styles, as mentioned above, is in the magma flux. Hawaiian eruptions occur both along fissure systems and from central vents. They tend to produce sustained fountains of lava clots that may reach several hundred meters above the vent or even higher in some cases. Lava clots or fragments in the inner cores of the fountains tend to retain their heat such that they coalesce upon deposition to form clastogenic lavas or densely welded spatter cones and ramparts (note, for example, that the Hawaiian-style activity at Pu'u O'o, Kilauea volcano, Hawaii, has generated 2.7 km^3 of lava by spatter coalescence since 1983; USGS, 2006). Pyroclasts around the outer edges of the fountains may cool and be brittle upon deposition. Head and Wilson (1989) summarize the proximal pyroclastic facies that are commonly produced by such eruptions (Table 1; see also Sumner, 1998). In the proximal cone area of deposition, relatively continuous layers of densely welded clasts form, often alternating with zones of moderate or weak welding caused by pulsing of the eruptive fountains (Fig. 1B; Head and Wilson, 1989; Valentine and Groves, 1996). The predominance of welding can result in a smaller component of grain avalanches and subsequent recycling of brittle clasts. Because of the sustained nature of the fountains and the potential for abundant cool clasts around the margins of the fountains, Hawaiian eruptions can produce tephra fallout deposits that extend several kilometers downwind from the vent (Parfitt, 1998). It is in these downwind deposits that extremely vesicular clasts (such as reticulite) may be found.

2.3. Violent Strombolian

We use the term violent Strombolian in the same manner as Valentine (1998) and Arrighi et al. (2001) to refer to explosive activity that produces sustained eruption columns up to ~10 km high and with the dominant clast sizes being ash to lapilli. Other authors have suggested that such behavior be referred to as sub-Plinian (e.g., Francis, 1993). However, it seems useful to have a term for the lower-altitude columns that we refer to as violent Strombolian. One reason for this is that many papers about sub-Plinian events refer to eruption columns ranging between 10 and 20 km altitude (e.g., Bursik, 1993; Ablay et al., 1995; Cioni et al., 2003; Vergniolle and Caplan-Auerbach, 2004; Arce et al., 2005; Coltelli et al., 2005; Sulpizio et al., 2005), although some authors use the term for columns with lower heights (e.g., Bonadonna et al., 2005; Folch and Felpeto, 2005) and Sparks et al. (1992) argued that the term sub-Plinian should only apply to eruption columns with heights less than 14–18 km. Francis et al. (1990) recognized the need for a term that would be tied to the low-altitude class of eruption columns and offered the term microplinian, and this term was used by Behncke et al. (2006) in their thorough account of recent activity at Southeast Crater (Mount Etna, Italy). We think that violent Strombolian is a better term, partly because it has become more familiar (although not well defined) since its introduction by MacDonald (1972), especially over the past decade as more attention has focused on explosive basaltic eruptions. One aspect that separates violent Strombolian, using our definition, from sub-Plinian, is that the eruptions of the former type do not penetrate the tropopause and therefore are likely to only have significant impacts over distances of several tens of kilometers from the vent, whereas the sub-Plinian eruptions will often reach into the stratosphere and therefore have potential impacts over much larger regions and even globally.

Violent Strombolian eruptions such as at Parícutin (see Luhr and Simkin, 1993) and Cerro Negro (Hill et al., 1998) appear to consist of multiple eruption column events, each with durations ranging from several minutes to several days. Coarse blocks and bombs fall out at low altitudes or, if ejected at an inclined angle, follow ballistic paths. Ash and lapilli that are carried upward in the convecting columns are dispersed downwind and can produce significant cm-thick accumulations up to a few tens of kilometers from the vent (see also Valentine et al., 2007).

Pyroclastic facies of violent Strombolian eruptions can be quite different from those of Strombolian and Hawaiian eruptions (Table 1), although the fact that the styles grade into one another (see Section 2.4) means that there can be significant overlap as well. Proximal (cone) facies of violent Strombolian deposits may have little to no welding because the well-fragmented clasts have adequate time to cool and solidify during their residence in the eruption column and subsequent fallout time. Clast shapes range from relatively sparse fluidal bombs to angular blocks and lapilli to subrounded shapes for clasts that have been recycled in the vent and re-ejected. A significant feature of violent Strombolian cone facies is that non-welded beds may be deposited by ash and lapilli fallout, simply mantling the cone slopes with variable components of grain avalanching (e.g., producing local lenticular reverse graded zones within beds that are otherwise laterally continuous as in Fig. 1C; Riedel et al., 2003; Valentine et al., 2005). This contrasts with classical Strombolian cone facies, in which avalanche features predominate, and with Hawaiian cone facies where some degree of welding is nearly ubiquitous. The large number of such beds in cone facies produced by violent Strombolian activity reflects a large number of eruption column events over a period of months to years during which a monogenetic volcano may be active (see Wood, 1980; Luhr and Simkin, 1993; Valentine et al., 2007). Many of these events are short-lived and produce only low columns that do not generate significant layers in the fallout deposits beyond the cone. An additional feature that is documented by Valentine et al. (2005, 2007) is the presence of localized, laminated and cross-laminated ash horizons. These may extend laterally for only tens of meters, and appear to record weak density currents that might have been produced by locally heavy ash fallout (e.g., Taddeucci et al., 2004; Valentine et al., 2007).

Medial to distal pyroclastic facies from violent Strombolian eruptions have bedding characteristics that are typical of fallout deposits, being planar-parallel and mantling topography. Textures within beds can range from massive to normal- or reverse-graded, depending upon plume evolution within a given eruption column event, and are clast supported. In proximal areas (within a few km from vent) the deposits commonly consist of fine to medium lapilli, and thin ash layers may cap some beds. At Lathrop Wells volcano (Valentine et al., 2007), many clasts in medial locations appear to be small ribbons or fragments of ribbons. Variable contents of abraded scoria lapilli are also present, recording recycling of clasts in the vent. At more distal locations (a few km to as much as ~20 km), deposits are mainly thin beds of ash (see, for example, Hill et al., 1998). Reconstruction of eruption column properties based upon the distribution of tephra deposits is complicated by the fact that a fallout blanket associated with a cone-forming eruption often records many eruption column events that occur over a period of weeks to years. Thus, an isopach or isopleth map for an entire fallout deposit associated with a scoria cone likely represents a composite result from many different combinations of column height and wind speed and direction (e.g., Self et al., 1974; Valentine et al., 2007). Rare cases where a single eruption column episode is observed (Hill et al., 1998) or where

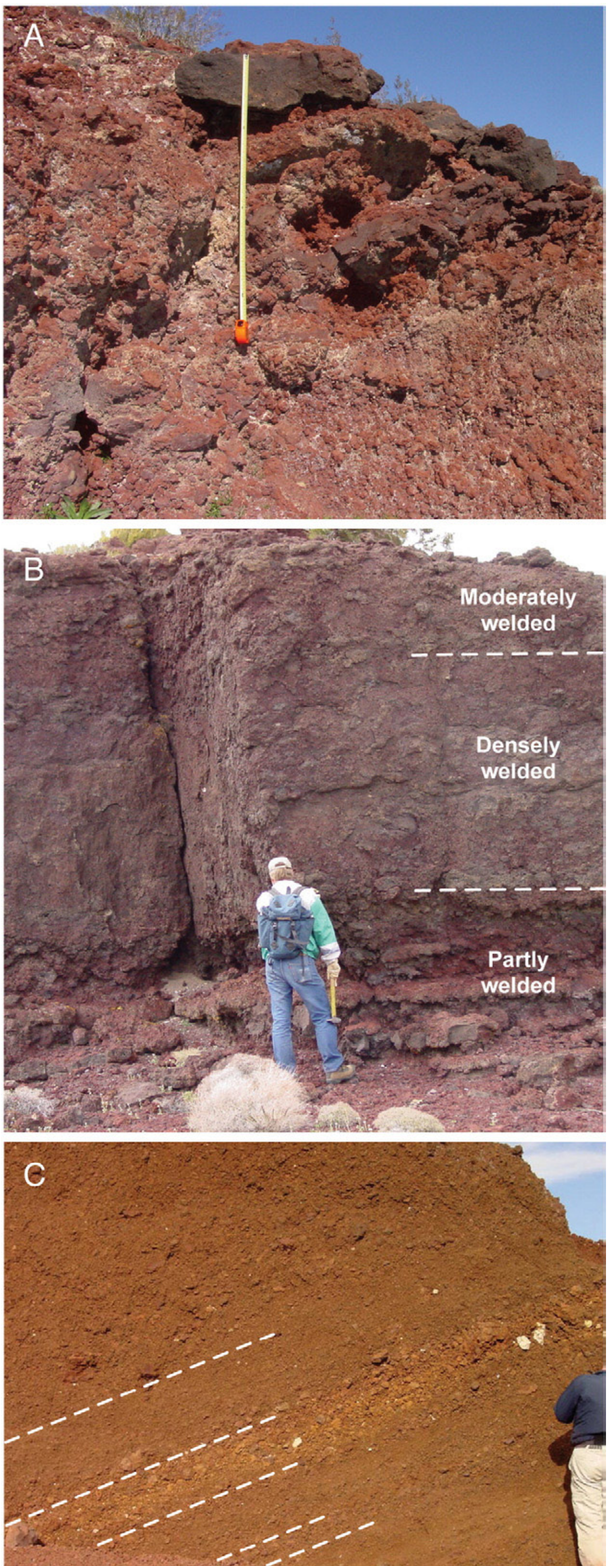

Fig. 1. Examples of proximal pyroclastic facies from three types of explosive basaltic eruption styles (all three examples are within ~100 m of vent). (A) Proximal Strombolian facies exposed at Black Cone, Nevada (~1 Ma; Valentine et al., 2006). Note abundance of coarse lapilli, which commonly occur in lenticular beds produced by grain avalanching. Local accumulations of coarser bombs (ranging from fluidal, flattened shapes to spheroidal) also define crude bedding. Measuring tape is 1 m long. Deposits are slightly welded, and dip toward the vent. (B) Proximal Hawaiian facies. These deposits are from an early phase of a Miocene-age fissure eruption at East Basalt Ridge, Nevada (Keating et al., 2008), deposited on flat ground about 50 m from the vent. Note partly welded basal deposits consisting of coarse lapilli and flattened bombs, grading upward into densely welded deposits (above person). Proximal (cone) facies of Hawaiian eruptions often have vertically alternating zones of moderate and dense welding, reflecting pulses in eruptive fountains. Person is ~1.8 m tall. (C) Proximal facies produced by violent Strombolian activity. Deposits consist almost entirely of medium to coarse lapilli and some blocks, in massive, planar-parallel beds (dipping toward vent, indicated by dashed lines) that are characteristic of deposition by fallout rather than avalanching (lighter colored, lithic-enriched bed is a good example of the bedding style). Deposits are non-welded, and fluidal or flattened bombs are rare. Person is ~1.8 m tall.

individual beds can be adequately traced and characterized over their extents (e.g., Arrighi et al., 2001) provide opportunities for reconstructing eruption column height.

2.4. Relationships between Strombolian, Hawaiian, and violent Strombolian

As mentioned above, the primary variables that determine the style of explosive activity for basaltic eruptions driven by magmatic volatiles are the mass flux and clast size (Fig. 2). Low mass flux, with discrete bursts of coarse juvenile bombs mixed with cool, recycled blocks and lapilli (see McGetchin et al., 1974) characterizes Strombolian activity. The coarse grain size of the bursts reduces the rate of heat transfer between particles and gas, so that a substantial buoyant ash column cannot develop. Although the bulk grain size distributions of Hawaiian eruptions are likely skewed toward the coarse part of the spectrum, the high, sustained flux can produce sufficient lapilli and ash to result in limited fallout deposits that extend some kilometers from the vent (Parfitt, 1998). Violent Strombolian eruptions may have high mass flux, like Hawaiian eruptions, but a much higher degree of fragmentation. This promotes effective heat transfer between clasts and gas and development of ash-rich eruption columns.

It is likely that for a mass flux in the range that can produce Hawaiian and violent Strombolian eruptions, the transition between the two can be defined by a critical grain size below which heat transfer is rapid relative to the time scale of the jet-thrust region of an eruptive column or fountain. The heat transfer time scale can be estimated simply as the thermal diffusion time scale (t_{diff}) for a particle of radius r, or

$$t_{\mathrm{diff}} \approx r^2/K, \tag{1}$$

where the thermal diffusivity (K) is assumed to have a value of 10^{-6} $\mathrm{m^2/s}$. This time scale can be related to the residence time in the jet-thrust region of an eruption column, which to first order can be determined by assuming that the jet-thrust height is simply the height at which the initial kinetic energy of the erupting mixture is spent (converted to potential energy) and vertical speed would decrease to zero if no buoyancy was added by air entrainment. The first order residence time for a clast in the jet-thrust region, t_{jet}, is therefore, ignoring drag between ambient atmosphere and the eruptive jet and any vertical changes in density of the erupting mixture,

$$t_{\mathrm{jet}} = v/g, \tag{2}$$

where v is the exit speed at the vent and g is the gravitational acceleration. A crude measure of the ability of an eruption jet to transform into an tephra-laden, buoyant eruption column is for a large fraction of the clasts to be sufficiently small that their heat is available to transfer to

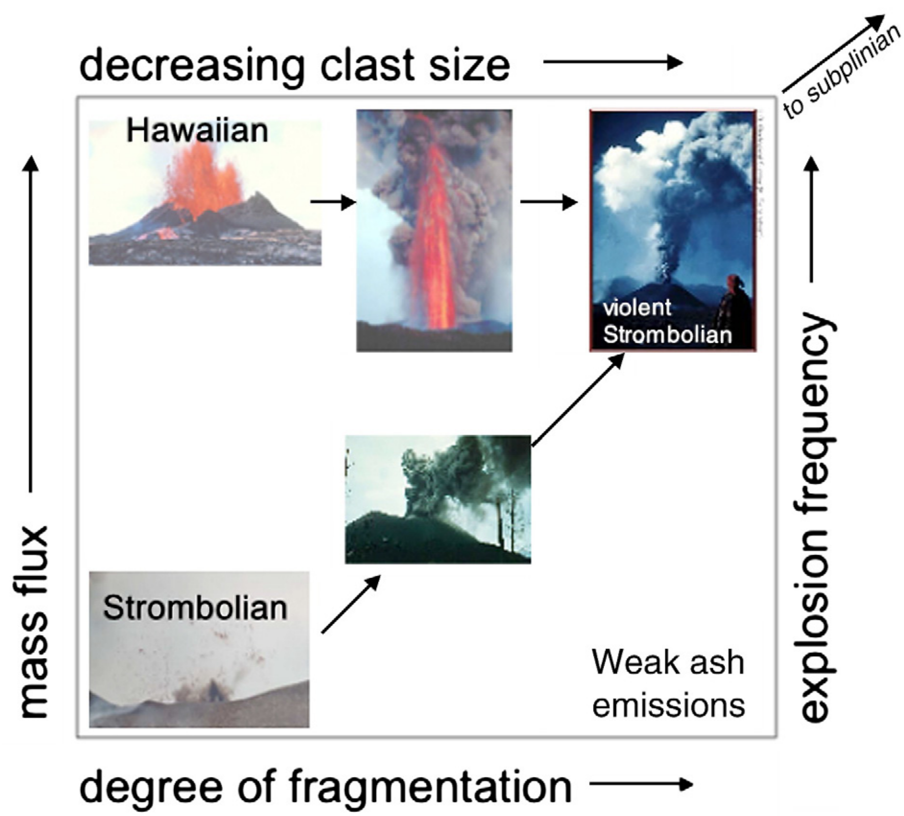

Fig. 2. Qualitative diagram showing styles of magmatic, basaltic explosive eruptions as functions of mass flux (and explosion frequency) and degree of fragmentation. Photographs of end-member examples of Strombolian, Hawaiian, and violent Strombolian (a.k.a. microplinian) eruptions are shown, as well as examples of Strombolian-violent Strombolian, and Hawaiian-violent Strombolian transitional behaviors. The Strombolian-violent Strombolian transitional example illustrates an eruption that is characterized by discrete bursts with a burst frequency that is nearing that necessary to produce an effectively steady plume, and although the bursts are coarse-grained, there is enough fine-grained tephra to produce some downwind deposits. The Hawaiian-violent Strombolian example has both coarse incandescent fountain component and a fine-grained buoyant plume component. With respect to the end-member facies characteristics discussed in Table 1, this eruption would produce both substantial downwind fallout deposits and extensive welded zones in proximal areas. Upper middle and lower left photos are from activity at Mount Etna in July, 2001 (obtained online at http://rivierc.club.fr/l'eruption_de_juillet_2001.htm, and used with permission), except for the Hawaiian example, which is the Pu'u O'o, Hawaii, activity of 6 September 1983 (courtesy of U.S. Department of Interior, U.S. Geological Survey). Center and upper right photos by K. Segerstrom. Note that increasing fragmentation and mass flux beyond the violent Strombolian (or microplinian) end member in this diagram will result in sub-Plinian to Plinian behavior as defined by Walker (1973a).

the gas phase within the jet-thrust region so that buoyancy can take over above that. This implies that,

$$t_{\rm diff} < t_{\rm jet}, \tag{3}$$

or,

$$r < (vK/g)^{1/2}. \tag{4}$$

For example, vent exit speeds of 50, 100, and 200 m/s, which likely are representative of violent Strombolian events, would require that a large fraction of clasts have $r<2$, 3, and 4 mm, respectively, by this crude estimate, to have clast-gas heat transfer rates sufficient to thermally couple the phases and promote formation of a tephra-laden, buoyant eruption column.

Therefore, a key question is, what fraction of clasts below a certain radius, for a given exit velocity, constitutes a sufficiently large fraction to drive the dynamics? If, for the purposes of the discussion in this paper, we apply Eq. (4) to the median grain size, we are in effect assuming that half of the erupted clasts need to be sufficiently small to satisfy Eq. (4) and that this will provide enough heat transfer to produce a buoyant column. The resulting clast size values are basically the same values estimated by Woods and Bursik (1991) for full thermal equilibrium. Eq. (4) would then imply that with sufficiently high mass flux, an eruption will be violent Strombolian rather than Hawaiian if the median grain size of the erupting mixture is in the fine lapilli range. Although bulk grain size distributions are difficult to obtain for prehistoric eruptions, this criterion is probably met by, for example, the deposits of the inferred violent Strombolian phase at Lathrop Wells volcano (see Valentine et al., 2007). This compares with the whole-deposit grain size distributions for the Hawaiian Kilauea Iki deposit, which has median clast diameters between ~64 and 256 mm (coarse lapilli to bomb sizes; Parfitt, 1998). Similarly, McGetchin et al. (1974) report an estimated median clast diameter of ~100 mm for the whole-deposit grain size distribution produced by Strombolian eruptions at Northeast Crater (Mount Etna, Italy) in June, 1969. A topic of future research would be to more rigorously determine what fraction of clast sizes must satisfy Eq. (4), rather than simply assuming that 50% is adequate.

The above discussion focuses on clast-gas thermal equilibrium within the jet-thrust part of an eruption column; Woods and Bursik (1991), on the other hand, focused on the effects of heat loss from a column due to fallout of clasts. They showed that eruptions with mean clast sizes larger than ~64–128 mm will have inefficient clast-gas heat transfer, and clasts will carry enough heat out of the column as they fall out of the jet-thrust region, so as to prevent the formation of a buoyant, convective column. Median clast diameters between those estimated with Eq. (4) and the 64–128 mm range will result in eruptive behaviors and pyroclastic facies that are transitional between Hawaiian and violent Strombolian end members (see below).

Explosion frequency, which is related to mass flux, also plays an important role in determining the character of an eruption (Fig. 2). Blackburn et al. (1976) defined Strombolian activity as "consisting of a series of discrete explosions separated by periods of less than 0.1 s to several hours" (frequencies up to 10 Hz and higher). Valentine et al. (2005) hypothesized that if the explosion frequency is greater than the characteristic rollover frequency of the largest-scale eddies at the base of an eruption column, the material in individual bursts will be mixed so that above some height the eruption column will be effectively steady. Assuming that the largest eddies at the base of a momentum-driven jet have radii similar to the vent radius (r_v), the rollover frequency (f) can be roughly estimated by

$$f \approx v/2\pi r_v. \tag{5}$$

For a vent radius of 5 m, exit velocities of 50,100, and 200 m/s yield $f\approx 2$, 3, 7 Hz. This simple heuristic argument is currently being tested with numerical experiments, but if true it would appear that the high-frequency explosions included by Blackburn et al. (1976) in their definition of Strombolian would imply activity that effectively produces steady eruption columns. If the high-frequency explosions are coarse-grained, such eruptions would be called Hawaiian by many workers (although the poor coupling between gas and coarse clasts might negate the eddy rollover argument), and if they are fine-grained such that a buoyant tephra column is able to form above the jet-thrust height many workers would identify them as violent Strombolian. We suggest that the original meaning of Strombolian, *a la* typical eruptions at Stromboli, should be adhered to such that the term only is applied to activity characterized by discrete bursts and discrete ejecta columns or fountains.

Parfitt (2004) also discussed the issue of explosion frequency with respect to eruption column dynamics. She pointed out that some eruption column events such as seen in 1973 at Heimaey (Iceland), where high-frequency explosions produced fountains of incandescent material up to ~250 m high with tephra-laden plumes up to ~6–10 km above them, also tend to simultaneously produce lava flows, a behavior significantly different from typical activity at Stromboli but more akin to Hawaiian eruptions. The simultaneous production of mafic lava and tephra-laden columns up to 10 km high seems to be common both at historical (e.g., Parícutin; see synthesis by Luhr and Simkin, 1993) and prehistoric eruptions (Valentine et al., 2007). Because of the combination of discrete (although closely spaced in time) explosions and simultaneous lava production, Parfitt (2004) referred to such eruptions as transitional between Strombolian and Hawaiian.

We suggest that there are different types of transitional behavior, based upon the combination of mass flux/explosion frequency and degree of fragmentation of the eruptive mixture (Fig. 2). For example, activity that is transitional between Strombolian and Hawaiian end members will have a low degree of fragmentation (clast sizes skewed toward the coarse lapilli and block/bomb size range); bursts of pyroclasts and gas might be temporally discrete, but with a frequency that is sufficiently high that proximal clasts from successive bursts are able to coalesce before solidifying and form clastogenic lava. Activity that is transitional between Hawaiian and violent Strombolian end members will have a high mass flux but might have an intermediate degree of fragmentation such that the eruption column has both an incandescent fountain of coarse clasts and a buoyant tephra column above it, along with simultaneous lava flows. Finally, activity that is transitional between Strombolian and violent Strombolian end members might also have an intermediate degree of fragmentation and bursts that are frequent enough that they verge on producing a steady, buoyant tephra column; fallout deposits from such activity might be characterized by many thin beds. Violent Strombolian eruptions within this context might also have simultaneous lava flows, but these are not clastogenic as in a Hawaiian eruption; rather, they tend to effuse from shallow breakouts (boccas) on the flank or at the base of a cone.

The mechanisms that might cause a basaltic explosive eruption to behave in the different ways are likely very complex and define important research topics. Parfitt (2004) summarized the role of magma rise speed and bubble separation in determining whether an eruption is Strombolian or Hawaiian. While this aspect is clearly important, other factors must come into play. For example, Valentine et al. (2005, 2007) showed that the Lathrop Wells volcano evolved from early Strombolian to late violent Strombolian styles, with lava effusion concurrent with both explosive styles and no significant change in major element composition. The volatile content of this trachybasaltic volcano was relatively high (e.g., H_2O in the range of 2–3 wt.% and possibly as high as 4.6 wt.%; Nicholis and Rutherford, 2004; Luhr and Housh, 2002), and there is no reason to infer that volatile content of the source magma changed through the eruption sequence, as there are only minor variations in major element compositions through the sequence. In fact, the trend toward more explosive activity later in the eruptive sequence would seem inconsistent with a decline in volatiles in the source magma that is commonly assumed in eruption sequences. Valentine et al. (2005) speculated that the change in eruptive style might be related to shallow degassing and microlite growth, which would both produce abundant bubble nucleation sites and increase the effective viscosity of the magma such that it behaved more akin to a silicic magma with respect to bubble and fragmentation dynamics; analyses are under way to test this hypothesis. However, the simultaneous effusion of lava from the same conduit complicates this view and suggests that two-phase flow phenomena with a high-speed mist surrounded by an annulus of liquid might be important as well. Cervantes and Wallace (2003) discuss the role of ascent rate and degassing on eruption style for the Xitle volcano in Mexico. It is clear that more studies of eruption facies, bubble and crystal textures and size distributions (e.g., Lautze and Houghton, 2007), and volatile contents are needed, along with theoretical and analog experimental studies (e.g., Seyfried and Freundt, 2000), to quantify the complex interplay of processes that can produce a wide range of explosive behaviors in basaltic volcanoes.

A final note on this section relates to Walker's (1973a) inference that violent Strombolian eruptions such as at Parícutin must involve explosive magma–water interaction. While this certainly is an important process in some cases and magma–water interaction is a topic that needs much more attention in the volcanological community, we suggest that violent Strombolian mechanisms do not *require* external water. For example, the Lathrop Wells volcano, except for one minor pyroclastic surge horizon of ambiguous origin, does not preserve any obvious record of hydrovolcanic explosions, but rather appears to have had sustained eruption columns driven by magmatic fragmentation (Valentine et al., 2007).

3. Lava flows

Lava flows are associated with both effusive (shields) and explosive (scoria cones) monogenetic volcanoes. The basaltic lava flows observed in these volcanic terrains display the full range of morphologies and inferred emplacement styles—and even some "new" emplacement styles that have not yet been observed while active. Here, we present our current understanding of how lava flows are erupted and emplaced at monogenetic basaltic centers, and point to areas that require additional investigations. To clarify the following discussion, we provide these definitions: an eruption is considered to cover the amount of time over which products emanate from a single vent. Because temporary pauses in eruptions are common (e.g., Wolfe et al., 1988; Mattox et al., 1993), multiple individual episodes of activity comprise a single eruption. As an example, the current eruption of Pu'u O'o on Kilauea volcano, Hawaii, began in 1983 and currently consists of 55 episodes (USGS, 2006).

3.1. Lava flow sources

Lava flows may erupt directly from the main vent of a cone, effusing quietly from an already open vent. This type of behavior may occur later in an eruption, after most of the volatiles in the system have already been released, but can also occur simultaneously with explosive eruption of an Hawaiian-style fire fountain (e.g., Kilauea Iki in 1959) or violent Strombolian style, as mentioned above. At cones where lavas effused directly out of the main vents, the pyroclastic cones are typically horseshoe shaped with the crater open on one side where lavas flow out. At the Pinacate volcanic field (Sonora, Mexico) many such craters open to the down slope side of cones that were built on sloping ground (Gutmann, 1979). Such lavas also may have rafted away significant masses of cone material (see Section 4). The open crater shape suggests that cone building was not able to keep pace with the rafting away of pyroclastic deposits atop the lavas. Other cones have horseshoe shapes due to strong winds that focused pyroclast deposition in one direction during cone building, rather than due to lava flow processes.

Lava flows also accumulate from lava-fountain fall-back (Head and Wilson, 1989; Sumner et al., 2005). To form such a clastogenic lava flow, spatter clasts must still be molten upon deposition on the ground and must accumulate at a sufficiently high rate that they coalesce before solidifying. Accumulated lava flows may form initially as a temporary pond within the summit crater of a scoria cone before breaking out as a flow. Sumner (1998) described an additional mechanism for generation of clastogenic lavas, where an initial spatter cone or rampart is formed that subsequently collapses after clasts weld under the lithostatic load, forming a lava flow but with some lag time after original pyroclastic deposition. Less intensely welded portions of the cone may be rafted along the back of the lava flow in this case (see Section 4). In some places, clastogenic lavas may preserve relict clast shapes and rheomorphic flow structures, but more commonly these flows are indistinguishable from lavas with purely effusive sources. The preservation of relict clast shapes must depend upon variables such as coalescence temperature, vesicularity, crystallinity, and shear; a topic of future research would be to quantify the conditions for preservation (or non-preservation) that might then be related back to eruption parameters (see Sumner et al., 2005).

In many cases, it can be difficult to determine the ultimate source of a lava flow associated with a scoria cone. Sunset Crater, for example (San Francisco Volcanic Field, Arizona) erupted two lava flows that today appear to have emanated from the base of the cone. If the flows were fed from a temporary lava pond within the crater of the main cone, there is no strong evidence for it. Lava boccas that might have sourced the lavas from the base of the cone were subsequently covered by late cone-building activity.

Several other historic and prehistoric cones fed lavas from their bases and provide clues about the controlling factors on bocca location. At Parícutin (see Luhr and Simkin, 1993), lava vents seem to be aligned along the top of a feeder dike, while at some scoria cone volcanoes of southern Nevada (Valentine et al., 2005, 2006, 2007), lava vent locations seem to have been mainly controlled by topography around the bases of the cones (lavas welled out at topographically low points). Quantification of the relative controls of feeder dike orientation, topography, and weaknesses within pyroclastic constructs is an important topic of future research. This could be particularly important in mitigating risks associated with formation of new volcanoes in basaltic fields (for example, in urbanized areas).

3.2. Lava flow types

Mafic lavas associated with continental monogenetic volcanoes display a full range of flow morphologies. Lava flow morphology is a complex product of viscosity (itself a product of temperature, crystal content, vesicle content, volatile content and composition), underlying slope, volumetric emplacement rate, shear rate, and cooling rate (e.g., Crisp and Baloga, 1990; Fink and Griffiths, 1990; Rowland and Walker, 1990; Gregg and Fink, 2000). The volume of the eruption is no indicator of the scale of processes. For example, the Roza member of the Columbia River Basalts is $\sim$1300 km^3 and is composed of individual pahoehoe-style lobes that are up to 50 m across (Self et al., 1997). The Wapi shield of the eastern Snake River Plain, in contrast, is $\sim$1.1 km^3, but contains pahoehoe-style lobes up to 30 m across (e.g., Greeley and King, 1977; Hughes et al., 2002b).

Flows associated with monogenetic shields, such as are common in the Snake River Plain, Idaho, are commonly fed by tubes or channels, and most abundantly display pahoehoe-style morphologies (e.g., Greeley and King, 1977). The largest lava tubes in the world (<100 km long) are associated with the Undara volcano (a monogenetic shield) in the McBride volcanic province of NE Australia (Atkinson and Atkinson, 1995). However, a'a type flow morphologies have also been identified in these provinces; similarly, both pahoehoe and a'a lava types have been found having erupted from scoria cones (e.g., Craters of the Moon National Monument, Idaho, USA).

Even a monogenetic structure may consist of many eruptive episodes, and the resulting lava flows may interfinger, anastamose, and pile on top of each other in a complicated way to generate a compound flow field. The area covered by a lava flow field associated with a monogenetic volcano depends on the total volume erupted, the lava viscosity, effusion rate, and pre-existing topography. A lava flow is considered to be cooling limited if it stops advancing because its crust has thickened to the point where the crust is strong enough to contain the flow; a lava flow is volume limited if it stops moving because there is no more supply (Chester et al., 1985). Entwined with cooling limited is viscosity: the colder a flow is to begin with, the higher its viscosity and the less area it is likely to cover. Similarly, the higher the crystal concentration within a lava, the more viscous (and probably colder) it is, and the more likely it is to be cooling limited.

Lavas that cover the greatest area are emplaced beneath an insulating crust, such as a pahoehoe flow or a tube-fed flow. The fractures in the surface of an advancing a'a flow allow the flow interior to cool relatively rapidly, making it difficult for it to achieve great lengths. There is an enormous body of literature discussing the differences and transitions between pahoehoe and 'a'a (e.g., Peterson and Tilling, 1980; Rowland and Walker, 1990; Cashman et al., 1999; and references therein), so we do not present details here. It is important to note, however, that the role of volatiles in the pahoehoe-a'a transition (e.g., pahoehoe may transition to a'a if sufficient volatiles are exsolved from the lava during flow; Peterson and Tilling, 1980) may suggest that a'a flows would be more commonly associated with scoria cones than with shields, and that the converse would be true for pahoehoe flows. To our knowledge, no such study has been undertaken.

In a pioneering study, Walker (1973b) correlated the lengths of lava flows with effusion rates. Malin (1980) showed that if tube-fed flows are included, the correlation no longer exists; Hawaiian tube-fed flows are generated by long-lived eruptions with a slow-but-steady (<3–5 m^3/s) effusion rate. A sudden increase in the effusion rate may cause the tube roof to rupture, generating a surface flow from the breakout; a sudden drop or long hiatus in the effusion rate can cause the tube to solidify (e.g., Mattox et al., 1993; Hon et al., 1994). Clearly, flow length is not a simple result of a single factor, but a complex interplay of many variables, although for some types of flows these other variables might have a secondary role compared to effusion rate.

Whether a tube, channel, or sheet flow is generated depends on the interplay of viscosity, effusion rate, and underlying slope (Fink and Griffiths, 1990; Gregg and Fink, 2000). High viscosity, steep slopes, and high effusion rates favor the emplacement of a'a channels, or toothpaste lava (Rowland and Walker, 1987); low slopes and low effusion rates favor pahoehoe lobes or tubes. However, Valentine et al. (2006, 2007) describe flows associated with small trachybasaltic scoria cones that were emplaced upon gentle slopes (~10–30 m/km) with apparently low effusion rates (based upon lava flow length per Walker, 1973b) that have surface textures most closely related to classic a'a. Partly because of this apparent contradiction, Valentine et al. did not apply the terms a'a or pahoehoe, which have many genetic connotations based upon better characterized, active lavas on Hawaii and Mt. Etna. Valentine et al. (2006, 2007) present evidence that the lava fields they studied also grew by a combination of stacking and inflation that have been described above, and that the lateral spreading of the lava fields to form fan-like constructs reflects an interplay between the gentle topography and local slope changes caused by lava flows themselves. It would appear that a quantitative understanding of the factors controlling emplacement of small volume lavas is an important area for future research.

Underlying topography plays a vital role in the emplacement of lava flows. As mentioned above, pre-existing topography is one control on where a flow may break out of a scoria cone, and topography certainly dictates whether a shield will be symmetric. The Undara lava tube system achieved its great length because the lava flows were confined within dry river beds (Atkinson and Atkinson, 1995). Similarly, the Roza member of the Columbia River Basalts was able to flow for several hundred kilometers from its source largely because it was confined to a canyon. Simply put, lava flows down hill, but details of lava flow are sensitive to any variation in topography that is equivalent to roughly 1/3 to 1/2 the flow thickness. In the case of inflated pahoehoe (Hon et al., 1994) an initial pahoehoe lobe may have a thickness of 30 cm and is therefore deflected and otherwise affected by features >15 cm. Later, the individual lobes may inflate to thicknesses of >5 m, but these lobes may still reflect the small-scale topography that influenced their earlier emplacement when they were thinner.

Investigations of lava flow morphologies on monogenetic shields within the Snake River Plain have revealed styles of lava flow emplacement that are not commonly observed on active basaltic systems today (e.g., Gregg et al., 2005). For example, Gregg et al. (2005) describe a form of emplacement of large (10's of meters across and 2 m thick) pahoehoe lobes that feed into each other by sequential inflation and deflation. The proximal lobe is emplaced, inflated, and then deflated—like a burst balloon—to feed the next distal lobe. More research is required to understand these unique emplacement mechanisms.

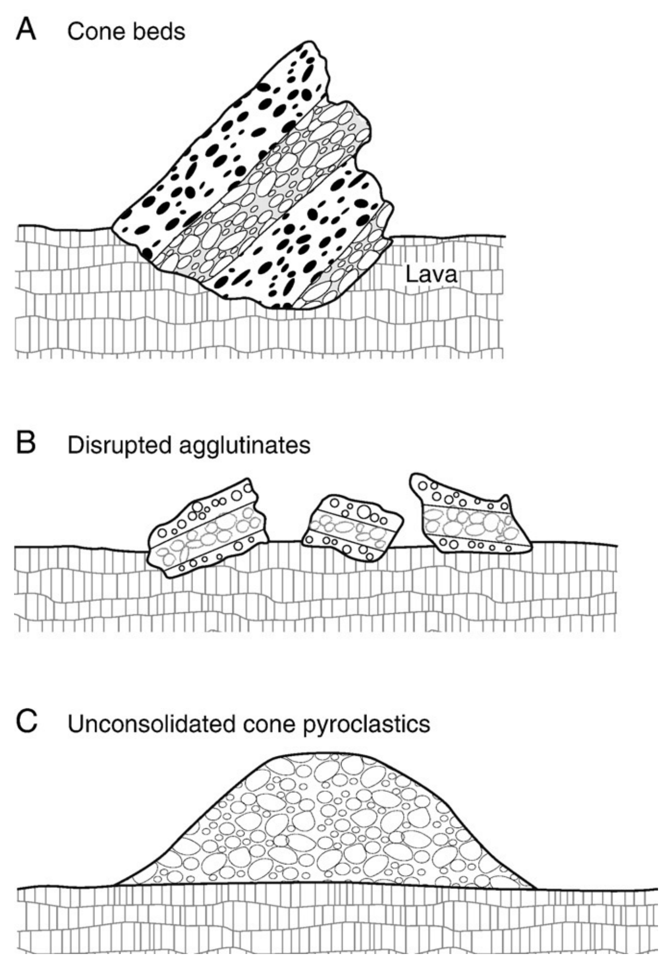

Fig. 3. Diagrammatic sketches of three types of rafts of proximal material atop lava flows. (A) Bedding of scoria and agglutinate, inherited from original deposition on the cone, remains intact. (B) Partly or completely welded pyroclastic deposits partly disaggregate during transport atop lava flow, resulting in large chunks of agglutinate. (C) Loose pyroclastic rubble with no relict of original bedding.

4. Interaction between lavas and cone building/destruction

At least two historical eruptions (Parícutin and Heimaey; see Luhr and Simkin, 1993, and Williams and Moore, 1976, respectively) have exhibited the process of rafting, where chunks or segments of proximal cone material collapses or is torn off the cone by departing lavas and carried atop the lavas to some location in the flow field. Prehistoric examples of rafted cone material on lava flow fields have been described at Sunset Crater (Arizona; Holm 1987), and at scoria cone volcanoes of the Southwestern Nevada Volcanic Field (Valentine et al., 2005, 2006, 2007). The authors have also observed such features at Craters of the Moon (Idaho, U.S.A.), Cima volcanic field (California, U.S.A.), and the Pinacate volcanic field (Sonora, Mexico). Valentine et al. (2006) identified three main types of rafts (Fig. 3): (1) those in which the stratigraphy of variably welded proximal beds remains intact (Fig. 4A), (2) rafts where welded or partly welded material partly disaggregates, resulting in large fragments of agglutinate that might internally preserve some bedding structure (Fig. 4B), and (3) rafts of loose, coarse pyroclastic material. Rafting is a common, but not ubiquitous, process that works against cone building. Some cones such as Parícutin (Luhr and Simkin, 1993), Sunset Crater (Holm, 1987), and Lathrop Wells (Valentine et al., 2007) retain no visible evidence of the loss of rafted material (for example, scarps or missing cone sectors) in the morphology of the cones themselves after eruptions have ceased. This indicates that cone-constructing pyroclastic activity outpaced the volume loss to rafting, although it is not necessary that rafting and healing are simultaneous. As mentioned in Section 3.1, cones that have "missing" quadrants (e.g., horseshoe shaped cones) might have been unable to keep pace with rafting. In such cases, if lava flow rates can be estimated, this information could be combined with an estimate of total volumes of rafted material atop the flow to calculate the rate at which the cone volume was lost to rafting. This in turn provides a constraint on the maximum rate of cone building. A topic of future research would be to survey a large number of cones and associated lavas to determine the range of cone volume loss rates to rafting as a function of cone facies, lava composition, effusion rate, and slope of the depositional surface.

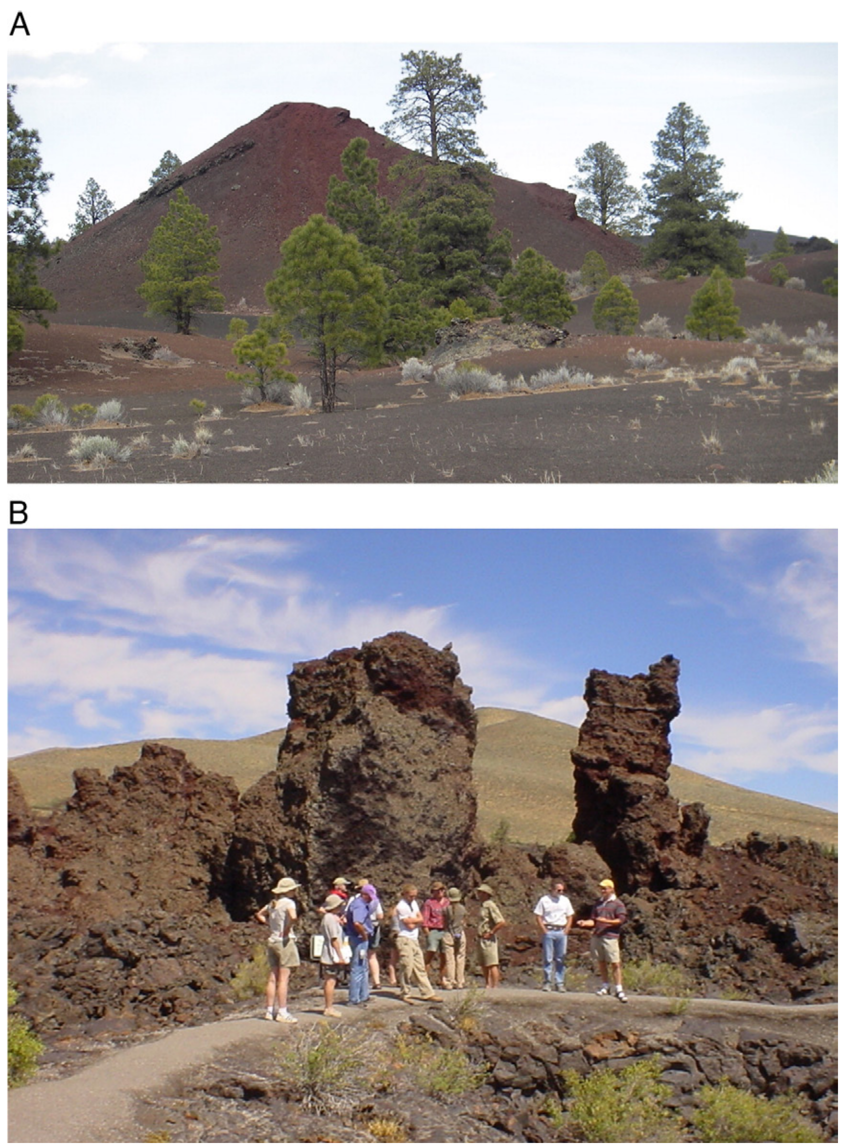

Fig. 4. (A) Raft of cone material on top of the Bonito flow at the ~1 ka Sunset Crater volcano (Arizona, USA), an example of the type of raft illustrated in Fig. 3A. Original bedding of variably welded cone deposits dip to the left and are locally visible through the scree that has formed due to raft degradation (see Valentine et al., 2006) as well as fallout scoria that blanketed the lava flow and raft surfaces. Raft is ~15 m high. (B) Rafted cone material at Craters of the Moon National Monument (Idaho, USA), illustrating the breaking up of cone agglutinate into blocks as the raft was carried atop a lava flow. This raft corresponds to the type illustrated in Fig. 3B.

Rafting suggests a range of cone behaviors with respect to lavas. Welded cone deposits might break away to form rafts by brittle failure or, if the deposits are sufficiently hot, in a plastic manner. The fragments of cone visible in the Craters of the Moon National Monument, for example, do not appear to have been deformed in a brittle manner, as in the formation of a crack or dike; rather, they have a distinctly ductile appearance. Unless the bulk density of a raft is lower than the carrier lava (which seems likely in situations where rafts are composed of relatively unwelded pyroclastic material), the lava flow must have sufficient yield strength to support the rafted fragments. The minimum yield strength therefore depends upon the mass of the cone fragment. Theoretically, critical yield strength represents the force that must be overcome for the flow to deform. In this case, one might argue that if the yield strength of the lava flow were too low, a dense cone fragment (a.k.a. if it was composed of densely welded proximal spatter deposits) would sink through the flow to the underlying ground, and would not be carried along the flow. Some rafts seem to have just begun this sinking process before coming to rest atop the lava; for example, Sumner (1998) and Valentine et al. (2006) describe lava squeeze-ups at the leading edges of rafts that record displacement of lava as it slowed to a stop under the weight of the rafts. There are many formulae to determine the yield strengths of lavas (e.g., Zimbelman, 1985); the result of interest to the rafting problem is the bulk behavior of the flow (including fluid interior and solid crust).

Finally, the flow must be able to break through the cone. If the flow is fed from a fire fountain that creates a lava pond in the cone's summit crater, than the pressure the lava pond exerts on the cone walls might exceed the strength of the cone and break through (a simple dam problem). If the flow is fed from a dike beneath the cone or a bocca at the cone base, then the cone is ruptured essentially from below. In this case, we envision that the lava is erupted, and a crust begins to form on the flow surface. With time, more lava is emplaced beneath the solid crust, generating an inflated flow (Hon et al., 1994). Continued inflation could cause the lava to uplift portions of the cone, causing the cone to become unstable. The lava yield strength would have to exceed the pressure exerted by the mass of the cone for inflation to occur beneath the cone; thus, once the cone ruptured, the flow would be capable of carrying away fragments. This process would likely carry away intact pieces of cone interior that have had a chance to partly weld, and would therefore retain original stratigraphy as it rafts away. Depending upon how rough the ride is atop the lava, such rafts might remain intact (Fig. 3A) or might partly disaggregate (Fig. 3B). The "ambient" state of stress within the cone probably plays a role in determining whether a given sector can be easily rafted. For example, Valentine et al. (2006) suggested that when a cone is emplaced upon a sloping surface, the downhill side of the cone will be in a relative state of tension and lavas emanating from that side will be most likely to raft material.

An additional process that might feed rafts is if the cone is growing (upward and outward) around a relatively stationary bocca; in such situations the departing lava might periodically take away chunks or piles of cone material that has been deposited on top of the bocca. This process is a likely origin for the type of raft shown in Fig. 3C. Quantification of the conditions that promote rafting and the interplay between cone construction and deconstruction by rafting is a topic that is worthy of more research.

5. Monogenetic shields

There is enormous variety in morphologies, volumes and compositions of monogenetic basaltic shields. Preliminary studies of morphologies within the Snake River Plain, for example, suggest two main types: typical "shield-shaped" shields, and shields with "hats" (Sakimoto et al., 2002; Hughes et al., 2002a, 2005). A shield with a "hat" displays a distinct break in slope near the summit, and the summit slopes are 2 to 3 times as steep as the flank slopes. A detailed geochemical and petrologic investigation of the lavas comprising the "hats" vs. the "shields" reveals that the "hats" are more volatile- and crystal-rich than the flanks. In some cases, the hats are revealed to be the same composition as the shields, but are composed of spatter rather than flows. Both scenarios suggest an increase in effective volatile content as the eruption progressed (Hughes et al., 2002a, 2005). Thus, these shields with hats are not simply degassed versions of scoria cones; rather, there are complicated magmatic plumbing issues revealed in their morphologies. An important issue to address with future research is the mechanisms for changes in eruptive processes associated with the hat-forming deposits relative to the rest of the parent shield. Other variations in shield morphologies are related to underlying slope and external volatiles. By definition, all shields share a roughly similar profile of 4°–11° flank slopes, but the variations in erupted volume and shield dimensions can span orders of magnitude.

6. Plumbing

In this section we focus mainly on the nature of magma feeder systems, or plumbing, of continental basaltic volcanoes in the uppermost ~1 km of crust. It is processes that occur at these shallow depths, such as dike-structure interaction, flow localization to form conduits, and secondary magma breakouts, that strongly influence eruptive processes. We also briefly touch upon the relationship between shallow plumbing and deeper processes such as the melt extraction zone in the mantle.

6.1. Dikes

Basaltic magmas ascend through most of the Earth's crust via self-propagating fluid driven fractures, or dikes (see Rubin, 1995 for a review). Part of the reason for this is that at depths below any significant degree of volatile exsolution (for H_2O this depth ranges from ~1.5–10 km for initial dissolved water contents of 1 and 5 wt.%, respectively), basaltic magmas have similar or higher density than typical continental crust host rocks. This precludes mechanisms such as diapir rise where the country rocks flow plastically around a less dense, rising blob of magma. Dike propagation in continental basaltic fields appears to be dominantly vertical, with lesser components of the lateral propagation that is often observed in rifts on the flanks of large volcanoes; this is corroborated by observations at fields that have a wide range of volume fluxes (e.g., Southwestern Nevada Volcanic Field — Valentine and Perry, 2006, 2007; San Rafael Swell — Delaney and Gartner, 1997; Snake River Plain — Kuntz et al., 2002). Upward dike propagation of basalts can be driven by either a high-pressure boundary condition at the base of a dike (e.g., a magma source region or crustal chamber) in situations where the dike remains connected to that pressure source (Fig. 5A), or if the "hydrostatic" pressure of magma at the base of a dike is less than the lithostatic pressure in the adjacent country rock (see discussion by Dahm, 2000a),

$$\int_{z_t}^{z_b} \rho_m(z) g dz < \int_{z_t}^{z_b} \rho_r(z) g dz, \tag{6}$$

where $\rho_m(z)$ is the density of the magmatic mixture as a function of depth z, and $\rho_r(z)$ is the density of country rock immediately adjacent to the dike (z_t and z_b are the depth of the dike top and bottom, respectively). The former (pressure boundary condition) case is likely to propel dikes upward from a melting region in the mantle. The latter case can be enhanced by volatile exsolution in the advancing tip of a dike where pressures can be significantly lower than the local lithostatic pressure (e.g., Rubin, 1995). Note that the criterion specified by Eq. (6) is not necessarily the same as a local density contrast between magma and rock at a given depth, which is commonly called upon as an upward driving mechanism for dikes. Rather, the criterion is

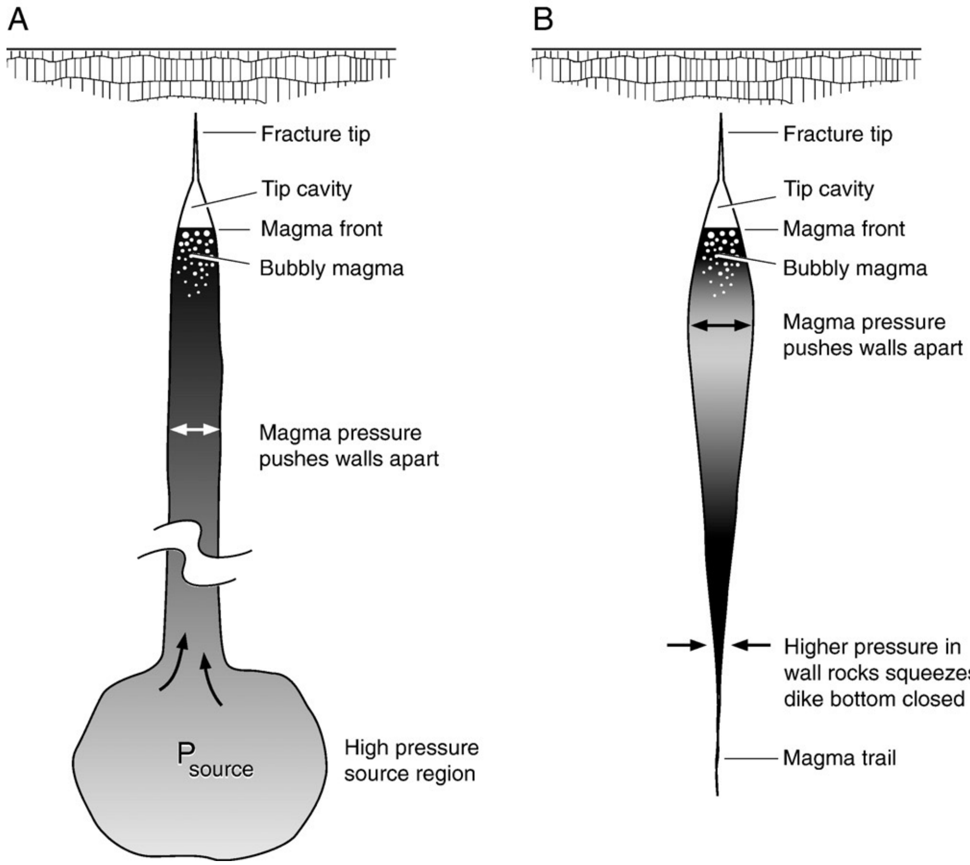

Fig. 5. Diagrammatic vertical cross sections illustrating two mechanisms for upward propagation of basaltic dikes through the crust. Diagrams are greatly exaggerated in the horizontal for illustrative purposes (real dikes would have depth to width ratios of ~10^3–10^4) (A) Dike remains connected to a high-pressure source such as the mantle source zone or a crustal magma chamber. (B) Dike separates from its source as it rises. Exsolution of volatiles into the dike tip cavity and in the bubbly magma front just beneath the advancing cavity may reduce the depth-integrated "hydrostatic" pressure at the base of the magma column relative to the lithostatic pressure in wall rocks at the same level, causing the wall rocks to squeeze inward and push the magma upward. Features in the area of the dike tip are reviewed by Rubin (1995). Lengths of dikes in the third dimension (perpendicular to cross section) are similar to the length scale of the magma's source region (see Section 6.5).

based upon the tendency of wall rocks to deform inward due to the pressure difference between the rocks and magma, thus pushing the dike closed at its base and squeezing the magma upward to drive continued propagation (Fig. 5B), while leaving a thin magma trail in the closed fracture beneath the dike (see Dahm, 2000b). Valentine (1993) suggested that upward dike propagation mechanism is enhanced because the low-pressure dike tip cavity causes vesiculation of part of the magma column at depths below the exsolution level based solely on lithostatic pressure, reducing the depth-integrated density of the magmatic mixture below that of the country rocks to satisfy Eq. (6).

At depth, and in the absence of structural heterogeneities such as faults, dike orientation is controlled by the orientation of the principal compressive stresses, such that the plane of a dike will always be perpendicular to the least principal stress (σ_3). If σ_3, which is commonly sub-horizontal, transitions to a vertical orientation, a dike may become horizontal (commonly called a sill, although the strict definition of a sill is that it is concordant with country rock bedding). Transitions in σ_3 orientation can be caused by dike-induced stresses on wall rocks (if the ambient differential stresses are small; Parsons and Thompson, 1991; Parsons et al., 1992), by lateral heterogeneities in country rock mechanical properties (particularly in faulted areas; Valentine and Krogh, 2006), and by local fluid pressure build-up when a dike is stalled at a rigidity contrast (see for example Kavanagh et al., 2006).

As a rising basaltic dike nears the Earth's surface it may be increasingly influenced by pre-existing structures. If a dike intersects a fault or joint whose plane is not oriented perpendicular to σ_3, the dike might be "captured" by the fracture if the magma pressure exceeds the compressive stress perpendicular to the fracture. At the scale of a volcanic field, this process may result in alignments of fissure vents and cones (e.g., Connor, 1990; Condit and Connor, 1996; Conway et al., 1997; Connor et al., 2000; Aranda-Gómez et al., 2003). On a smaller scale, fault capture of dikes affects the geometry and location of different features of the plumbing of individual volcanoes. Delaney et al. (1986), Baer et al. (1994), Delaney and Gartner (1997), and Ziv et al. (2000) discuss examples where dikes occupy pre-existing joints, and Valentine and Krogh (2006) and Keating et al. (2008) describe cases where dikes beneath scoria cone volcanoes and fissure-fed eruptions clearly occupied pre-existing normal faults. When a dike is captured by a fault in an area where the state of stress is near failure, the loss of friction across the fault, as it fills with liquid magma, can result in co-intrusive fault motion (e.g., Parsons et al., 2006; Gaffney et al., 2007). Valentine and Krogh (2006) show an example of how the interplay between dike intrusion, fault motion, and country rock heterogeneity influenced the dimensions of dikes and the localized formation of shallow sills beneath a scoria cone volcano. The complex, transient stress fields associated with ascending dike tips can also activate motion on pre-existing faults (e.g., Rubin and Pollard, 1988; Rubin, 1992) and can cause formation of tension fractures on the surface (e.g., Kuntz et al., 2002).

There is an important feedback between dike injection and tectonics (see discussions by Bursik and Sieh, 1989; Parsons and Thompson, 1991; Connor and Conway, 2000). In areas that are undergoing extension (e.g., Basin and Range Province of North America, or the African-Arabian Rift system), which is a common type of setting for basaltic fields, injection of dikes into the shallow crust can take up some or all of the tectonic strain. Thus, in an area with high magma flux as reflected by high vent density and large volumes of lavas, there may be little faulting even though the rate of extension is high. An example of this is the Snake River Plain in the northwestern U.S.A., where Basin-and-Range faulting is greatly reduced, even absent, across the volcanic field even though such faulting is very active on either side of the field (e.g., Parsons and Thompson, 1991; Kuntz et al., 2002). In a case such as the Snake River Plain, dikes are therefore not likely to be captured by normal faults in the uppermost crust and the orientation of eruptive fissures (the intersection of dikes with the Earth's surface) will be aligned perpendicular to σ_3 (see Parsons and Thompson, 1991, for examples). Conversely, in low magma flux fields such as the Southwestern Nevada Volcanic Field where there are few dikes to take up active extension, normal faults are abundant, dikes are more susceptible to fault capture, and eruptive fissures might be oriented according to the structural grain rather than the principal stress orientation at the time of volcanism. Aranda-Gómez et al. (2003) discuss an example of a single volcanic field where parts have relatively low flux and the location of vents are strongly tied to faults, and where other parts have relatively high magma flux and there is little relationship between vent location and faults. An important topic of research is quantification of the relationships between extension rates, faulting, and magma flux (on both the scale of a volcanic field and of an individual volcano) — which is likely to have a bit of a “chicken or egg” component given that the generation of mafic magmas in the mantle may be variably coupled to extension in the shallow crust.

6.2. Dike to conduit transition

Many, if not most, continental basaltic eruptions begin as fissure eruptions that represent the initial intersection of a dike with the Earth's surface. The flow of magma commonly focuses into distinct conduits that represent local widening of the feeder dike system as eruptions progress. Many different mechanisms for this widening have been proposed (reviewed by Valentine and Groves, 1996; Keating et al., 2008), including both mechanical and thermal erosion processes. From a modeling perspective, only two end members — dike-fed (fissure) flow and circular conduit flow — are treated in the literature (e.g., Wilson and Head, 1981; Papale, 1998; Mastin and Ghiorso, 2000; Mastin, 2002). A conclusive theoretical treatment of the transition between the two is lacking, probably hampered by the wide range of processes that are involved. For example, Bruce and Huppert (1989, 1990) consider just the role of thermal erosion (melt back) in modeling the transition from fissure to focused flow, whereas Delaney and Pollard (1981) focus also on mechanical brecciation processes. The issue of what generally controls the initiation and location of conduits is a topic that is ripe for further quantitative study, both in the field and in development of dynamical models.

6.3. Conduit geometries

Although there are several published field studies of the geometry of hydrovolcanic conduits (diatremes; e.g., Lorenz, 1986; White, 1991; Németh et al., 2001), we are aware of only one paper (Keating et al., 2008) that addresses conduit geometry for monogenetic basaltic volcanoes that were driven entirely by magmatic volatiles. Keating et al. describe five sites in the western U.S.A. where basaltic volcanoes have been sufficiently eroded to expose the upper ~200 m of their plumbing, but still have remnants of eruptive deposits that allow at least crude reconstruction of eruption styles. The eruption styles include Hawaiian fountaining from fissures up to ~3.5 km long, with 3–5 vents representing areas of focused flow, to Strombolian cone-forming eruptions from a single vent and associated lava flows. They show that at four of the sites, dikes that are typically 4–12 m wide at depths >100 m gradually branch out and widen upward. Above 50–70 m depth the dikes transition into elongate conduits that flare dramatically to form vent areas up to ~110 m wide (e.g., Fig. 6). Dike widening and gradation into conduits is a reflection of several processes, including wall-rock brecciation and entrainment, branching of individual dikes into pre-existing fractures to form a dike swarm, and inelastic deformation of wall rocks. Keating et al. compare these observations with simple steady-state, one-dimensional models of conduit flow (e.g., Wilson and Head, 1981; Mastin and Ghiorso, 2000), and show that although the absolute values of natural conduit diameters are much larger than those in theoretical calculations, the shapes of natural and theoretical conduits match very well when the theoretical conduit flow is assumed to be balanced with local lithostatic pressure (rather than the other end-member approach to conduit modeling, which assumes a constant diameter; e.g., Papale, 1998). It is inferred that the difference in absolute values of conduit radius at a given depth reflects the fact that the theoretical shape represents a single steady-state flow event, while the natural conduits preserve the superimposed shapes of many magma flow events over the lifetime of a volcano (probably on the order of months to years). In other words, the entire conduit as preserved in the geologic record was not active at any given time during the life of the volcano, but only channels or sub-conduits that migrated within the larger footprint over time. The similarity in natural and theoretical shapes validates the arguments of Wilson and Head (1981) and Mitchell (2005) that basaltic eruptions will evolve rapidly towards pressure-balanced conditions in the conduit due to failure of wall rocks under any significant pressure differences. However, many more studies at eroded volcanoes are necessary to fully understand the range of mechanisms and the generality of the conclusions of Keating et al. (2008). Of particular interest would be constraints on the rate at which the pressure-balanced conduit shape is obtained for a given flow event, and how this would affect changes in eruptive style in early phases of an event that might be reflected in the pyroclastic facies.

6.4. Processes in the conduit and effects on eruption style

In Section 2.4 we briefly discussed some of the issues of conduit dynamics that require further research to understand controls on eruption styles and transitions between them. These include shallow degassing, groundmass crystallization, and changes in mass flux. In addition to the studies of conduit geometry discussed above, it would be useful to explore whether there is any record of these processes in conduits and dikes that are exposed at eroded sites. Of course, it must be realized that exposed conduits and dikes reflect a combination of multiple eruptive events and are strongly overprinted by the very final stages of magma flow; any inferences about specific eruptive processes must filter out these aspects. Evidence for variations in mass flux, for example, might include nested chilled margins or breccia zones. Degassing might be preserved as geochemical variations in wall rocks (e.g., WoldeGabriel et al., 1999) around the conduits.

6.5. Relationships between shallow and deep plumbing

Each basaltic volcano ultimately represents liquid tapped from some volume of its partially molten mantle source region, and an intriguing problem is to try and understand whether there are links between physical characteristics of the volcano (e.g., volume, eruption mechanisms), its crustal feeder system, and the mantle source zone. Valentine and Perry (2006) discussed the concept of the “magmatic footprint,” which relates the length scale of feeder dikes with a characteristic length scale of the mantle source zone tapped by each volcano. It appears that these length scales are very similar in the volcanic field that they described — in other words, dikes are about as long as the characteristic diameter of the region of mantle that is tapped to feed a volcano of a given volume, and together these define the magmatic footprint (note that this argument might not hold in

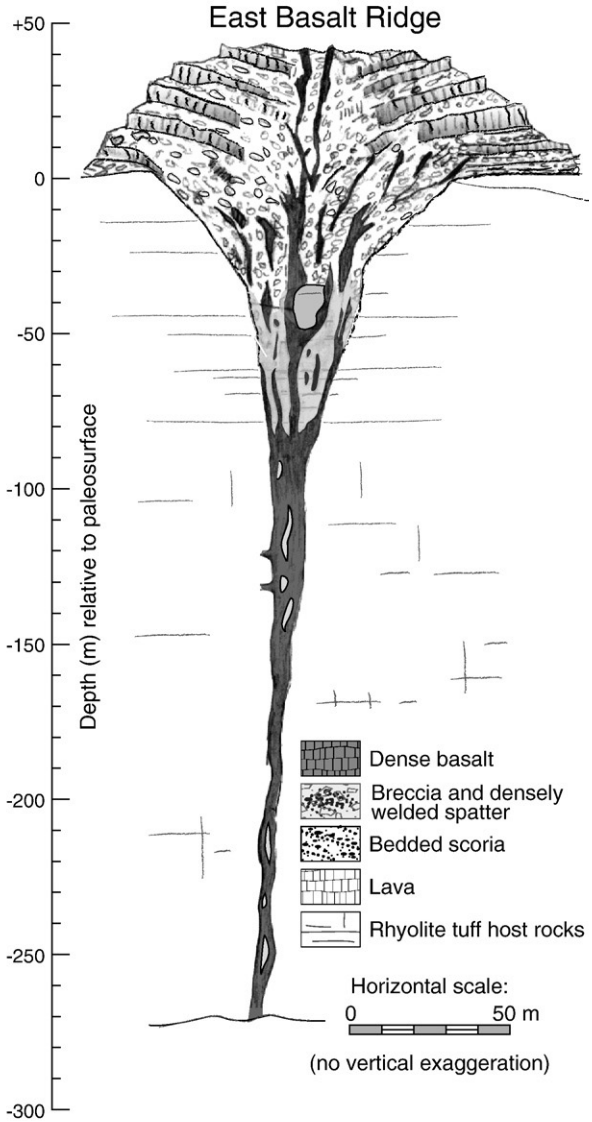

Fig. 6. Sketch illustrating shallow geometry of feeder dike and vent of a small basaltic volcano (~8.8 Ma East Basalt Ridge, Nevada, USA), showing shallow flaring of vent/conduit and vent facies deposits. Modified from Keating et al. (2008).

volcanic fields were there are substantial crustal magma reservoirs). The location of a monogenetic volcano is determined primarily by the location of the mantle source zone, and by features such as topography (Gaffney and Damjanac, 2006) and pre-existing faults (Valentine and Keating, 2007) that occur within the magmatic footprint. Implicit in their conclusion is that dike propagation is primarily upward through the crust rather than laterally, which is consistent with studies mentioned above. For example, a 4-km-long dike will likely not migrate toward a topographic depression or fault that is 10 km away.

If studies of other continental basaltic fields support the notion that the length scale of near surface features such as shallow feeder dikes and eruptive fissures are similar to the length scales of magma source zones, there are important ramifications. Detailed studies of compositional variations at monogenetic basaltic volcanoes in the Basin and Range province, such as the Southwestern Nevada Volcanic Field (see Bradshaw and Smith, 1994; Perry et al., 1998; Valentine et al., 2006, 2007) and the Cascade volcanic arc (Strong and Wolff, 2003), are demonstrating that there are subtle compositional variations within individual volcanoes that can only be explained by heterogeneity of the mantle source (see also Schmincke, 2007, for a discussion of similar mantle source heterogeneity related data from the Eifel volcanic fields in Germany). Valentine and Perry (2006) suggested that studies of these source-related compositional variations at fields where volcanoes are undergoing systematic changes in their magmatic footprints (e.g., at the Southwestern Nevada Volcanic Field, where magma flux is waning, magmatic footprints have systematically decreased in size by factors of ~5–10 over the past ~5 Myr) could provide information on the statistical properties of spatial variations in the underlying lithospheric mantle. This is a variation on the mantle "chromatography" approach that looks at variations in the composition of upwelling asthenospheric mantle beneath, for example, Hawaiian volcanoes (e.g., DePaolo, 1996). To understand how such source variations are preserved in eruptive deposits, it will be necessary to theoretically constrain the mechanisms acting between source and eruption that might smear or mix

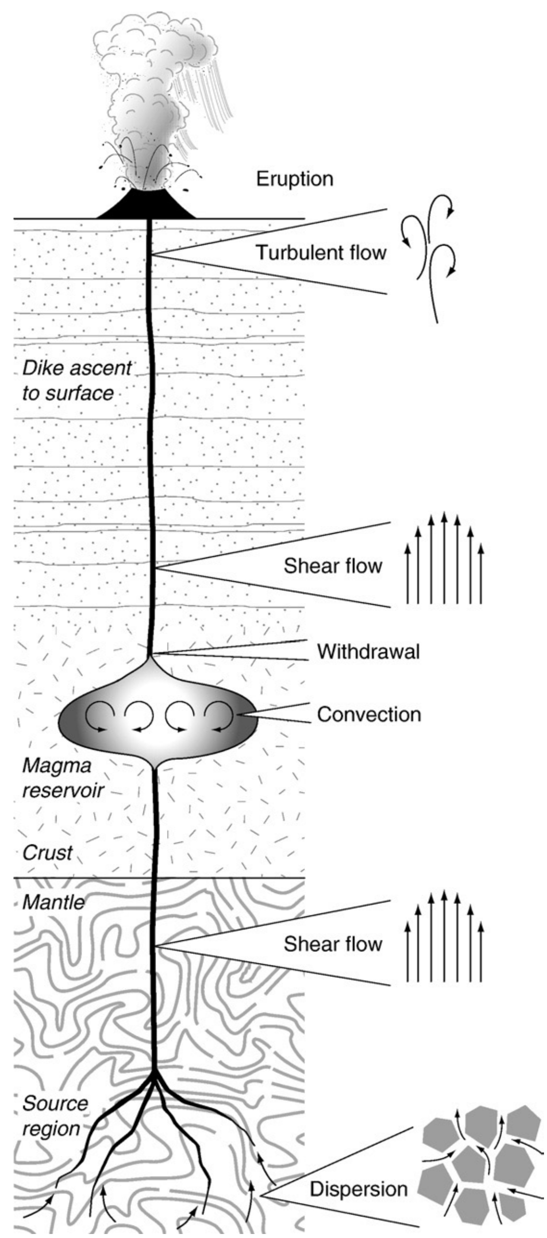

Fig. 7. Diagram illustrating fluid dynamic processes that might affect original source-related compositional variations in the melt, at different steps in the ascent of basaltic magmas.

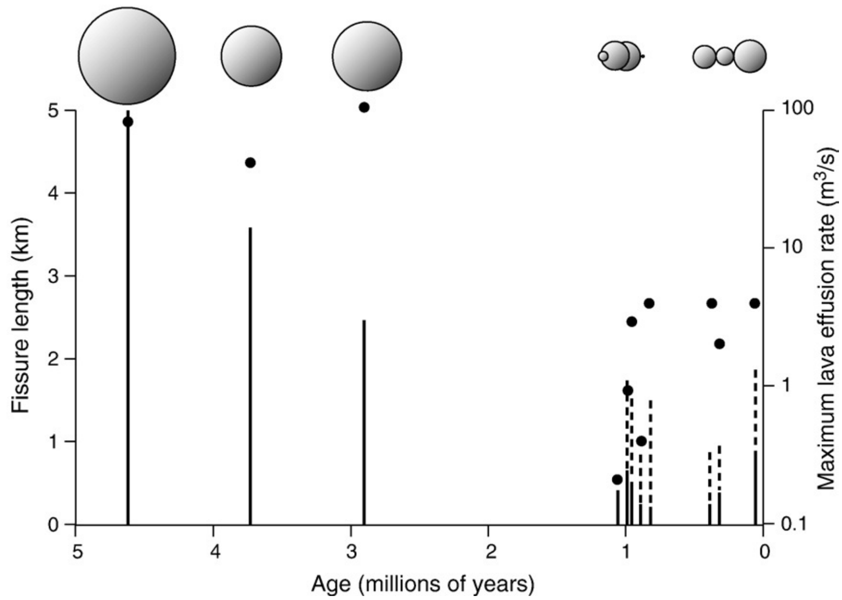

Fig. 8. Plot of eruptive volume (circles sized in proportion to cube-root of volume), fissure lengths (lines), and lava effusion rates (dots) for basaltic volcanoes in the Southwestern Nevada volcanic field during the past five million years (modified from Valentine and Perry, 2006).

those variations. These include dispersion as melt migrates through tortuous porous media paths in the melt production area, shear during dike ascent, convection during storage in crustal magma chambers, and turbulence as the magmas accelerate through the upper crust and erupt (Fig. 7). Mantle chromatography in continental basaltic fields may provide much information on the geostatistics of the uppermost mantle as well as providing constraints on the types of fluid dynamic processes that dominate magma ascent.

Table 2
Summary of problems for further research in physical processes associated with continental basaltic volcanoes

Topic	Issue	Possible Approaches
Explosive eruption styles	How bulk particle size distribution determines transitions from Hawaiian to violent Strombolian end members of activity.	Numerical and analog experiments of effective heat and momentum coupling between gas and ensembles of particles.
Explosive eruption styles	Role of explosion frequency in transition between Strombolian and violent Strombolian end members of activity.	Numerical and analog experiments of pulsed jets.
Explosive eruption styles	Interplay between vesiculation, groundmass crystallization, and degassing in explosion style.	Crystal- and vesicle-size distribution of products of different styles, experimental determination of viscosity as a function of groundmass crystallization, modeling of degassing processes coupled to magma flow.
Lavas	Relationship between degree of preservation of relict clast shapes in clastogenic lavas and eruption conditions.	Numerical and analog experiments of accumulation conditions, shear deformation, and welding.
Lava source	Quantification of the relative controls of feeder dike orientation, topography, and weaknesses within pyroclastic constructs on location of lava vents.	Field observations of variably eroded cones, analog experiments.
Lava flow types	Quantification of role of volatiles in determining pahoehoe versus a'a.	Field and petrologic studies comparing lava textures to estimated volatile contents.
Lava flow types	Dynamics of small volume lava flows.	Field studies of emplacement features and analog experiments.
Interaction between cones and lavas	Quantification of range of cone volume loss rates to rafting as a function of cone facies, lava composition, effusion rate, and slope of the depositional surface.	Field survey of a large number of cones and associated lavas.
Shields	Origin of hat-shaped shields.	Field, geochemical, and petrographic studies of products of different parts of hat-shaped shields.
Dikes	Role of dike tip vesiculation in ascent as determined by multiphase flow in dikes and elastic deformation of walls.	Numerical and analog experiments.
Dikes	Quantification of transitions between fault-dominated and dike-dominated extension in basaltic fields.	Chronological and geomorphic studies of basaltic fields with ranges of magma flux and/or tectonic extension. Analog experiments.
Dike to conduit transition	General predictive theory of location and rate of dike to conduit transition.	Field and geophysical studies of active and eroded vents, analog and numerical modeling.
Conduit geometry	Test whether flared conduit shape (pressure-balanced conduit flow) is general across volcanoes and volcanic fields.	Field studies of eroded vents and conduits, and quantification of shallow xenoliths in eruptive facies.
Relationships between shallow and deep plumbing	Test whether magmatic footprint concept applies to basaltic fields generally.	Field determinations of fissure lengths, eruptive volumes, effusion rates, and estimates of mantle source volume.
Relationships between shallow and deep plumbing	Determination of mantle compositional heterogeneity at different scales.	Quantification of source-related compositional variations within individual volcanoes with different magmatic footprints, and mixing processes during magma ascent.
Relationships between shallow and deep plumbing	Conditions that cause a basaltic field to transition to having more evolved polygenetic volcanoes.	Field studies of fields with and without polygenetic volcanoes in similar tectonic settings. Theoretical studies of heat and mass transfer effects with distributed dikes.
Relationships between shallow and deep plumbing	Mechanisms that relate effusion rate to volume and magmatic footprint at individual volcanoes.	Theoretical and analog experimental studies.

Some basaltic fields consist only of scattered monogenetic basaltic volcanoes, whereas others such as the San Francisco Volcanic Field of northern Arizona (U.S.A.) produce major polygenetic, andesitic to rhyolitic volcanoes. Intuitively, this must be related to the basaltic magma and heat flux into the crust as well as the state of stress and the composition of the crust. An important topic of research is to quantify the conditions that allow a field to produce multiple magma and volcano types through time, rather than only basaltic products. The concept of a magmatic footprint might aid in constraining the dimensions of basaltic feeder dikes that transport heat and mass into the crust from below, as a function of overall volume flux of a field.

Finally, the fundamental mechanisms underlying the linkage of shallow and deep length scales and volume fluxes need to be quantified. Valentine and Perry (2006) showed that relatively large magmatic footprints of individual volcanoes correspond with large effusion rates. As the length scale of eruptive fissures and the mantle source decrease by a factor of 5–10, effusion rates decreased by two orders of magnitude in the Southwestern Nevada Volcanic Field (Fig. 8). The relationships between these observations, as well as the level of generality with respect to other basaltic fields, need further research (see Valentine and Perry, 2007).

7. Summary

Continental basaltic volcanoes encompass a broad array of explosive and eruptive processes. In terms of eruptions driven by magmatic volatiles, explosive styles include Hawaiian, Strombolian, and violent Strombolian and gradations between them. A set of facies characteristics has been defined for the three end-member explosive behaviors. Any given basaltic volcano may exhibit one to all three of the styles at various times in its short monogenetic lifetime, and these variations are not always clearly related to factors such as variations in volatile contents that are commonly called upon. Rather, processes such as compositional variations, degassing, ascent rate, and multiphase flow must all be treated in a coupled manner to understand the wide range of behaviors. Effusive processes appear to encompass the same styles as are observed at active volcanoes such as on the island of Hawaii and at Mount Etna, but also appear to include lava flow dynamics that have not yet been well quantified. Lava flows and pyroclastic cone building are coupled at many volcanoes through the process of rafting. Finally, the ascent of basaltic magmas through the shallow crust includes complex interactions with pre-existing structures and topography that are also tied to the relationships between overall dike injection rates and tectonic strain. Characteristics of shallow plumbing provide insights into deeper source plumbing for both individual volcanoes and for basaltic volcanic fields.

Further studies of the physical processes we have discussed are important in order to further our understanding of the most common type of continental volcanism and the fundamental processes of basaltic magmas. Table 2 summarizes the problems that we have suggested as important future research directions. Addressing these problems will in turn further our ability to predict and mitigate the risks associated with basaltic volcanism, which threaten millions of people around the world.

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