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Chapter 9 Deposits of Hydroclastic Eruptions

Many volcanic eruptions result from the interaction of magma and external water (Table 9-1), but few volcanologists (e.g., Jaggar, 1949) have emphasized the importance of nonmagmatic water in volcanic eruptions. In our view, the importance of external water in explosive eruptions is still underestimated. Wood (personal communication) even holds that maars, which most commonly develop from hydroclastic eruptions, are the second most common volcanic landform on earth next to scoria cones.

The recognition of the influence of external water on volcanic eruptions has opened a Pandora's box of new problems that offer a broad field for future studies. To name only a few: How do (1) the geometry and size of vent, (2) magmatic parameters such as chemical composition, temperature, viscosity, and difference between solidus and liquidus temperatures, and (3) the amount of external water, as well as reservoir properties such as porosity and permeability, interact to produce steam explosions? What is the mixing mechanism that causes what appears to be virtually instantaneous incorporation of country rock, thorough fragmentation and vent-coring? How can the relative roles of magmatic and external volatiles be estimated in deposits that contain vesiculated tephra indicating at least some exsolution of magmatic volatiles? Can the claim of some authors be verified that many explosive eruptions are actually favored by the interaction of magma and external water? We have outlined present ideas on the main processes believed to govern hydroclastic eruptions in Chapter 4. Here we will concentrate on the deposits.

Definition of Terms

We use the name *hydroclastic* (introduced by Fitch in Walker and Blake, 1966), almost in parallel with pyroclastic, as an inclusive term for products of *hydroexplosions* which are defined as "explosions due to steam from any kind of water" (Wentworth, 1938, p. 22). Many types of explosively produced hydroclastic particles are expelled through vents, however, and therefore are *sensu stricto* pyroclastic, but if it can be determined that they are derived from steam explosions, the term hydroclastic is more appropriate than pyroclastic. A major reason that we generalize hydroclastic is that some workers have equated the terms phreatic and phreatomagmatic (Gary et al., 1974) and have used them for nearly any kind of explosion caused by the interaction of magma or lava with external water (e.g. Daly, 1933, p. 311).

Table 9-1. Different ways that hydroclastic eruptions can occur. (After Schmincke, 1977a)

1. Magma erupted into shallow ocean or lake waters
2. Water flowing into an open vent containing magma
3. Ascending magma intersecting a ground water horizon
4. Water flowing into a partially emptied magma chamber
5. Magma erupted beneath a glacier
6. Lava or hot pyroclastic flows travelling over wet ground or entering a body of water
7. Groundwater heated by, but not mixed with, magma

Table 9-2. Hydroclastic surface eruptions and corresponding attributes of deposits. (After Schmincke, 1977a)

Characteristics of particles and deposits		Eruptive and transport processes
Chemical composition: predominantly basaltic	indicates →	Low magmatic volatile content, high temperature, low viscosity
Clasts: Slightly vesicular, Sideromelane, Bread crust/cauliflower bombs	indicates →	Quenching and granulation at magma-water contact; minor degassing; steam explosions
Grain size: Small. May contain large lithic clasts and broken pillows	indicates →	Thorough breakage by thermal stress and absence of separation of fine clasts in eruptive column
Sorting: Poor	indicates →	Abundant water (vapor) in system
Structure: Penecontemporaneous deformation, Vesicular tuff, Good bedding, Mud cracks, Accretionary lapilli (vesicular), Deposition on vertical planes, Poor sorting	indicates →	Abundant water (vapor) in transporting system
Abundance of lithic clasts in deposits:	indicates →	Sudden steam generation within country rock and spalling
Large size of lithic and essential clasts:	indicates →	High energy due to abundant water (vapor)
Abundant evidence for horizontal transport:	indicates →	Base surge transport
Geometry: Low to very low angle cone	indicates →	Ballistic and base surge transport
Fumarolic alteration or pipes: Absent	indicates →	Deposition at low temperatures
Sintering or welding: Absent	indicates →	Low temperature (< 100 °C)
Association: Strombolian deposits	indicates →	Fluctuating supply of external water or sealing of conduit walls

Stearns and Macdonald (1946, p. 16–17) divide hydroclastic (hydro-explosion) eruptions into four categories: (1) *Phreatic eruptions* (explosions) are driven by the conversion of ground water to steam. Such steam explosions have a low temperature and do not expel juvenile ejecta because ground water is vaporized only by heat or hot gases and not by direct contact with fresh magma. More recently, Muffler et al. (1971) introduced the term "hydrothermal explosion" for steam explosions that do not produce fresh magmatic tephra, but occur when ground water is heated by an igneous source and flashes to steam which violently disrupts the near-surface confining rocks. We consider the terms hydrothermal eruptions and phreatic eruptions to be synonymous. (2) *Phreatomagmatic eruptions* (explosions) (Stearns and Vaksvik, 1935) occur when ascending magma contacts ground water; resulting eruption products include juvenile as well as cognate ejecta. Limiting such eruptions to ground water environments, however, is too restrictive. Here, we generalize phreatomagmatic to include magma and water interactions within any environment – submarine and sublacustrine as well as ground water, ice and wet-sediment environments. They differ from magmatic explosions, which are caused by internal magmatic gases.

As originally defined (Stearns and Vaksvik, 1935), phreatomagmatic refers to magma interacting only with ground water. Therefore, Walker and Croasdale (1972) proposed the name "Surtseyan" for eruptions of magma (commonly basaltic) through sea water. Schmincke (1977a) interpreted the deposits of the 1888–1890 eruption of Vulcano (Italy) as phreatomagmatic and therefore suggested that the term Vulcanian be used synonymously with phreatomagmatic (Chap. 4). (3) *Submarine explosions* occur when magma rises into the shallow sea, producing abundant juvenile glassy fragments. Sublacustrine and subglacial explosive activity give rise to similar kinds of ejecta; therefore we prefer the more inclusive term, *subaqueous,* rather than submarine, and regard explosive submarine eruptions as one variety of phreatomagmatic eruptions. (4) *Littoral explosions* (Stearns and Clark, 1930) occur where subaerial lava flows or hot pyroclastic flows meet water. As with underwater explosions, littoral explosions are regarded here as a variety of phreatomagmatic eruptions.

Self and Sparks (1978) defined "Phreatoplinian deposits", within Walker's (1973) classification, as formed by the interaction of water and silicic magma. Such deposits differ from Plinian deposits principally by a higher degree of fragmentation.

Rittmann (1958; 1962) introduced the term *hyaloclastite* for rocks composed of sideromelane clasts produced by essentially nonexplosive spalling and granulation of rinds of pillow lavas by increase in diameter of pillow lava tubes during growth. Since then, the term has been expanded to include vitric tuff from shallow-water explosive volcanism (Tazieff, 1972) as well as sideromelane-bearing tuff produced by lava flowing into water (Fisher, 1968a) and occurring in maar volcanoes (Fisher and Waters, 1970; Heiken, 1971, 1972, 1974; Schmincke, 1974a). Thus, we use the term to include all vitroclastic tephra produced by the interaction of water and hot magma or lava whether or not the interaction is associated with venting. The general characteristics and origin of hydroclastic deposits are given in Table 9-2.

Fine-grained material believed to have formed by nonexplosive granulation that is commonly associated with pillow lavas has been called *aquagene tuff* by Carlisle (1963). Honnorez and Kirst (1975), emphasizing the need to distinguish between the clastic products of nonexplosive and explosive origin, introduce the term *hyalotuff*, a glassy *pyroclastic* rock resulting from phreatic or phreatomagmatic explosion.

Components of Hydroclastic Deposits

Grain Size Distribution

Deposits from hydroclastic eruptions are characteristically fine-grained, although coarse-grained lapilli- and tuff-breccias are common in some deposits. Grain size analyses of phreatomagmatic deposits have been reported by Waters and Fisher (1971), Sheridan (1971), Sheridan and Updike (1975), Crowe and Fisher (1973), Schmincke et al. (1973), Yokoyama and Tokunaga (1978), Nairn (1979) and Self et al. (1980). Data on tuff ring deposits are summarized by Walker and Croasdale (1972) and Walker (1973). Walker (1973), in a study of 88 samples from tuff rings in the Azores and Iceland, shows that the median diameter is less than 1 mm in

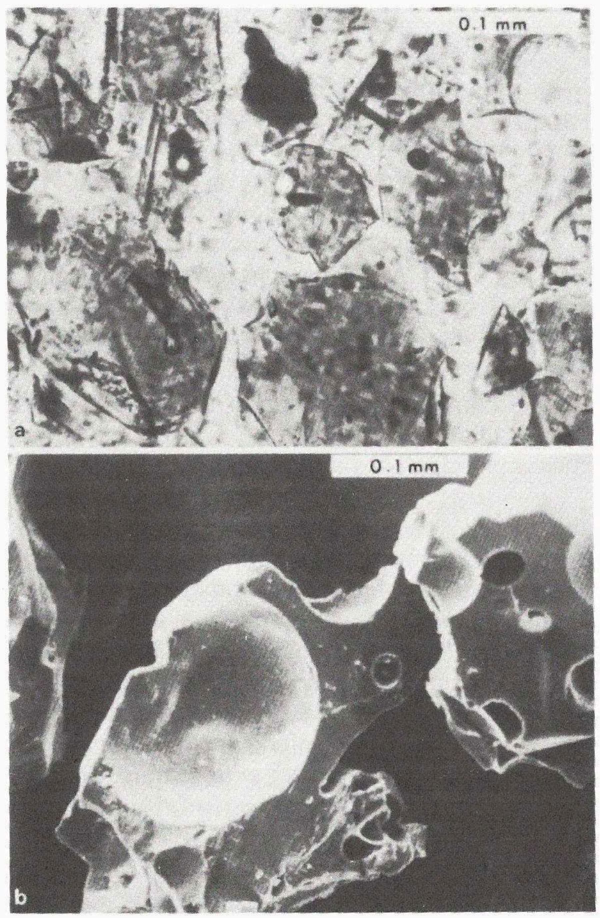

Fig. 9-1 a, b. Photomicrograph of hydroclastic shards. **a** Rectangular and polygonal mafic shards with notable scarcity of vesicles (Heiken, 1974; pl. 24 B), **b** Blocky mafic shards showing breakage across vesicles (SEM photo from Heiken, 1974, pl. 24 A). Shards are from Surtsey (Iceland)

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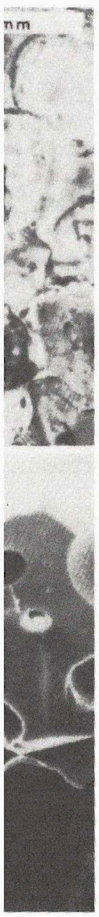

Fig. 9-1 a, b. Photomicrograph of hydroclastic shards. **a** Rectangular and polygonal mafic shards with notable scarcity of vesicles (Heiken, 1974; pl. 24 B), **b** Blocky mafic shards showing breakage across vesicles (SEM photo from Heiken, 1974, pl. 24 A). Shards are from Surtsey (Iceland)

about 75 percent of the samples. Qualitative inspection of many hydroclastic volcanoes composed wholly or partly of phreatomagmatic deposits leads us to suspect that there are essentially two main groups with many gradations in between. One group, represented by the ash and tuff cones, probably results from shallow explosions. Deposits of this group appear to be finer-grained and much better sorted than those of a second group represented by many maars which result from more powerful eruptions. The median diameters and sorting coefficients of the second group occupy a field between most flow and fallout deposits (Schmincke et al., 1973).

Characteristics of Essential Components

Vitric shards from hydroclastic eruptions are mostly mafic but silicic varieties also occur (Heiken, 1972, 1974). Most characteristic are blocky, nearly equant shapes with fracture-bounded surfaces transsecting few vesicles (Fig. 9-1). Some vitric hydroclastic shards have abundant mosaic cracks (Fig. 9-2) indicating rapid chilling.

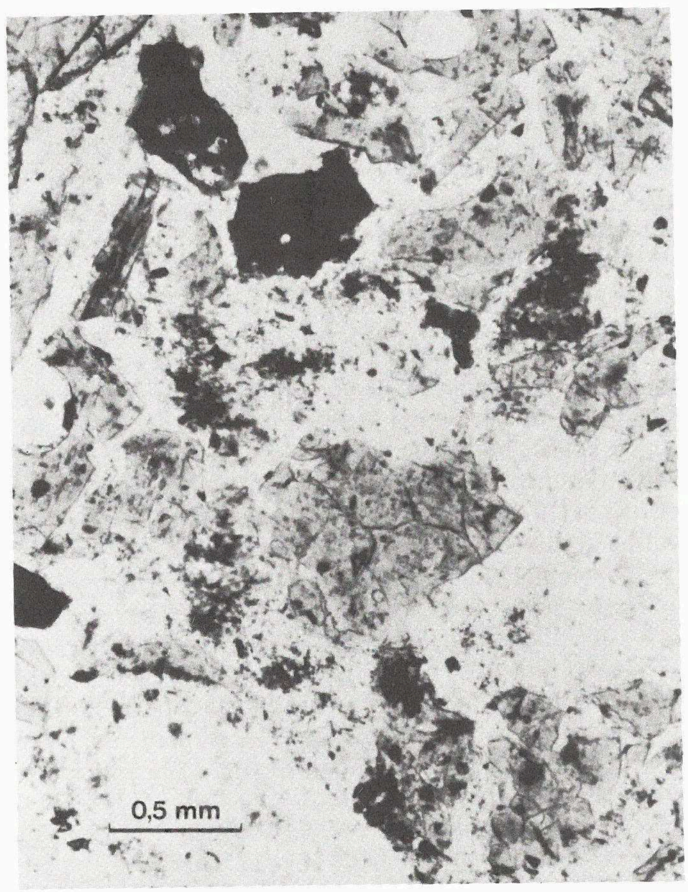

Fig. 9-2. Equant sideromelane grains showing jigsaw cracks. Pahvant Butte, Utah

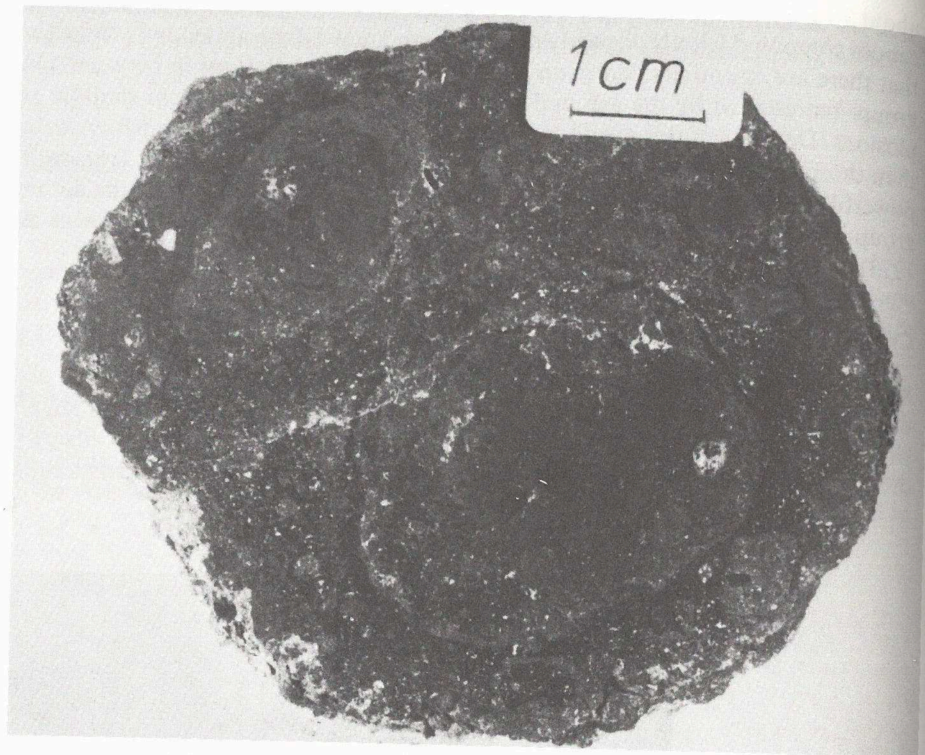

Fig. 9-3. Composite lapillus consisting of two larger and many smaller lapilli set in a dense lava matrix. Deposits formed during transition Vulcanian-Strombolian type eruptions. Prehistoric Marteles Crater (Gran Canaria, Canary Islands) (Schmincke, 1977a)

Blocky mafic shards are common in deposits of maars and tuff rings (Fisher and Waters, 1970; Waters and Fisher, 1971; Walker and Croasdale, 1972), and littoral cones (Fisher, 1968a). Those that are common in seamounts and within the interstices of subaqueously chilled pillow basalt are described in Chapter 10. Honnorez and Kirst (1975) have stressed that blocky, nonvesicular sideromelane shards form during deep water eruptions (below the critical depth of magmatic volatile exsolution) by granulation of extruding lava in addition to spalling of pillow lava rinds. So far, no deposits of significant volume have been described to illustrate this process.

Shards formed by cracking due to thermal shock are typically glassy and nonvesicular. Indeed, these features are a major argument for steam explosions as the main eruption mechanism leading to the formation of maars and tuff rings. However, there are many tuffs associated with maar and tuff ring deposits which are made of shards that are both vitric and slightly to highly vesicular. These shards are apparently formed by a combination of vesiculating magma and quenching by water or steam. Deposits made of such shards – which may show all transitions from blocky through slightly vesicular with scalloped edges to highly vesicular – are characteristic of shallow water eruptions. They may be the most common type

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Fig. 9-4. Composite lapilli (cf. Fig. 9-3) formed during transitional stage between Vulcanian (maar-forming) and Strombolian (cone-forming) eruptions. Marteles Caldera (Gran Canaria, Canary Islands) (Schmincke, 1977a)

of ash produced under water and typically occur in seamounts and in the transition from seamount to oceanic island.

Dense to slightly vesicular subspherical lapilli, some consisting of smaller lapilli held together by a lava matrix, occur in many cinder cones and maars within deposits transitional between phreatomagmatic and Strombolian deposits (Schmincke, 1977a) (Figs. 9-3 and 9-4). Layers made largely of such composite lapilli characteristically are inversely graded, caused by rolling down the slopes of the cones. Composite lapilli are similar in size, shape, and structure to "autoliths" described from some kimberlite diatreme breccias and believed by some workers to have formed in the presence of water (Lorenz, 1975; Schmincke, 1977a). Their subspherical shape and internal structure suggest formation and solidification within a vent prior to extrusion. They are interpreted to form when lava droplets are ejected into steam above the level of a magma column; the droplets are quenched and fall back to acquire a rind of new lava, and the process may be repeated several times.

Nakamura and Krämer (1970) first pointed out that surfaces of many lapilli and bombs from hydroclastic eruptions are characterized by a peculiar texture variously described as cauliflower, crackled or bread crust. This texture somewhat resembles that of bread crust bombs (Fig. 9-5), but unlike breadcrust bombs

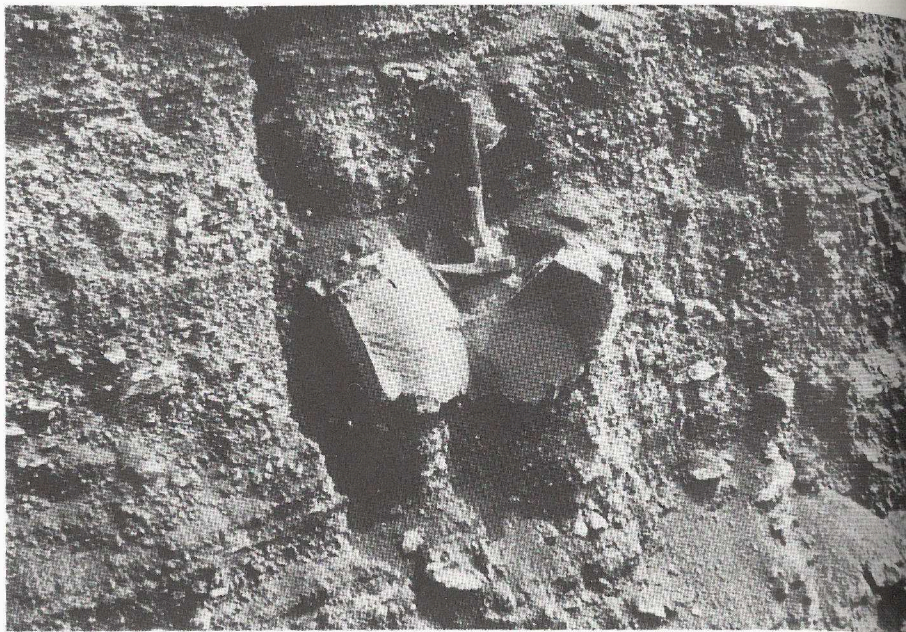
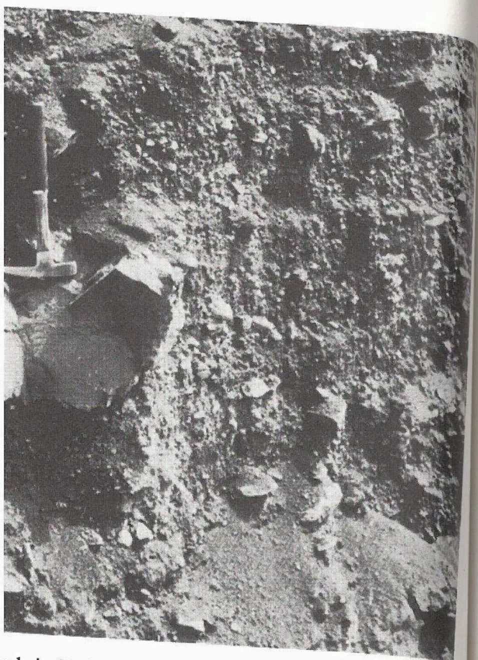


Fig. 9-5. Dense, quenched basanite cauliflower bomb in Vulcanian deposits rich in accidental clasts. Lummerfeld Maar (East Eifel, Germany) (Schmincke, 1977 a)

formed by internal gas expansion, cauliflower bombs commonly have dense or only slightly vesicular interiors (Schmincke, 1977a). Greatly expanded bread crust bombs, such as those formed during the historic eruptions of Vulcano (Walker, 1969), are not found among basaltic ejecta from hydroclastic eruptions because interior residual gas pressures of basaltic bombs are relatively low.

Accretionary Lapilli

Accretionary lapilli, which occur in many fine-grained ash layers, were reported from the 1965 phreatomagmatic eruptions at Taal Volcano (Moore et al., 1966), and they have been recognized in many subsequent investigations of hydroclastic deposits (e.g. Fisher and Waters, 1970; Heiken, 1971; Swanson and Christiansen, 1973; Lorenz, 1973, 1974; Schmincke et al., 1973; Self and Sparks, 1978). The occurrence of accretionary lapilli is not conclusive evidence for hydroclastic eruptions, however, because they are common in fine-grained fallout deposits where moisture is supplied by rain that often accompanies pyroclastic eruptions (Moore and Peck, 1962). The abundance of accretionary lapilli in hydroclastic tephra may be due to three factors: (1) abundance of water and steam in the eruption column, (2) production of abundant fine-grained tephra in hydroclastic eruptions, and (3) base surge transport, leading to deposition of fine-grained particles close to the source in contrast to Plinian eruptions where most fine-grained particles are usually deposited far from the vent, out of range of moisture related to the eruption column. One feature of some hydroclastic accretionary lapilli not described from



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other kinds of accretionary lapilli is the occurrence of vesicles in their outer layer (Lorenz, 1974) and in their core (Schmincke, 1977a).

Armored lapilli (Waters and Fisher, 1971) are a variety of accretionary lapilli containing crystal- or rock-fragment nuclei coated by rinds of fine to coarse ash. They range in diameter from 3 or 4 mm to as much as 10 cm or more, depending to some extent on the size of the nucleus, and have been reported only from hydroclastic deposits. In some cases, flattened lapilli- to bomb-size debris composed entirely of ash without cores are observed in hydroclastic deposits; these were apparently sticky and wet balls of ash when deposited. Armored lapilli apparently develop because the ash cloud contains abundant cohesive ash that sticks to solid particles within it. This mechanism differs from that proposed by Moore and Peck (1962) where moisture or rain drops falling through dry ash in eruption clouds causes agglutination of ash particles. In hydroclastic eruptions, there are probably all transitions between eruptions where large volumes of nearly pure water are initially ejected (Nairn et al., 1979) to blobs of wet ash to individual ash particles coated with moisture within vapor-rich eruption clouds.

Accidental Clasts

The form and shape of accidental clasts depends on the type of country rock at the site of fragmentation. Sandstone clasts, for example, are usually angular and blocky, whereas slate clasts are naturally platy. Accidental clasts commonly show little or no signs of thermal metamorphism, suggesting that temperatures are relatively low in much of the hydroclastic eruptive system (Fig. 9-6). Accidental clasts also occur as inclusions in essential lapilli and bombs; most such clasts are only slightly metamorphosed (Schmincke, 1977a) (Fig. 9-7). This suggests that fragmentation and incorporation of the country rock into the magma occurred shortly before or during eruption and thermal quenching. Some maar deposits contain abundant cobbles and pebbles derived from alluvial gravels of an underlying aquifer. Indeed, the presence of such material provides suggestive evidence for steam explosions within buried alluvial gravels, although it is possible that gravel could fall from surface levels into the zone of explosions.

Accidental rock fragments in maar and tuff ring deposits provide additional important information: (1) their maximum size allows estimation of explosion energy; (2) the type of crustal rocks present permits inferences about explosion depths if crustal stratigraphy is known; and (3) mantle-derived ultramafic xenoliths are common in some maar deposits.

Maximum Size of Fragments Related to Energetics

In the Nanwaksjiak Maar, Alaska (200 m deep; 600×1000 m in diameter), maximum diameters of blocks decrease from 3.3 m at the rim to 70 cm at 2000 m from the rim. From size relationships, Rohlof (1969; see also McGetchin and Ullrich, 1973) estimated that the eruptive fluid in Nanwaksjiak Maar had a density of about 0.01 g cm^3 and a surface velocity of about 500 m/s assuming an ejection angle of 65°. At Hole-in-the-Ground Maar, Oregon, Lorenz (1970) reports similar maximum diameters around the rim, but compared to Nanwaksjiak, the decrease in size away from the rim is less pronounced (Fig. 9-8). Lorenz calculated that pres-

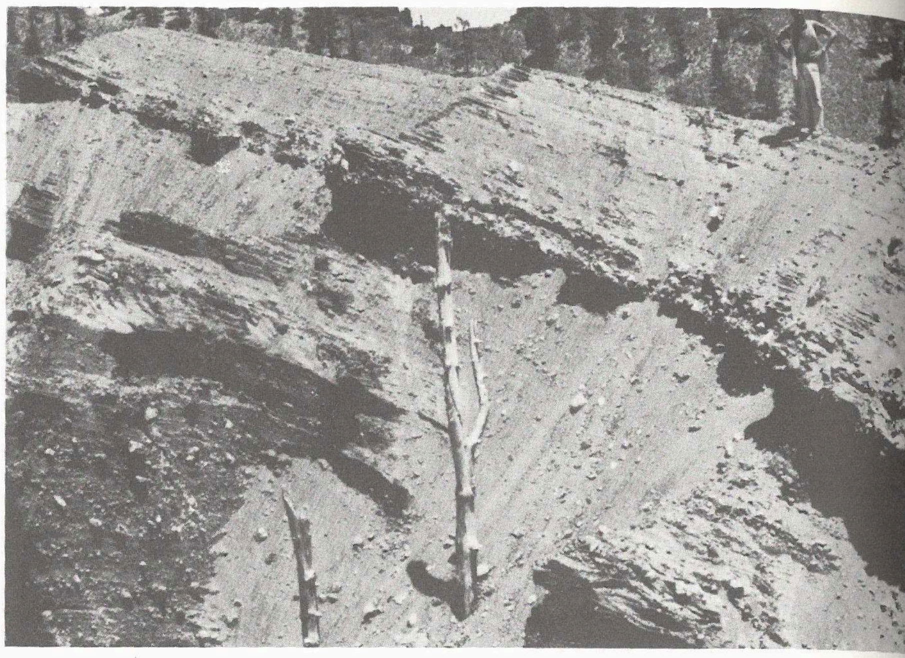


Fig. 9-6. Xenolith-rich base surge deposits from 1949 eruption of Duraznero Crater (La Palma, Canary Islands). Note debarked but not burned pine trees rooted on pre-eruption surface

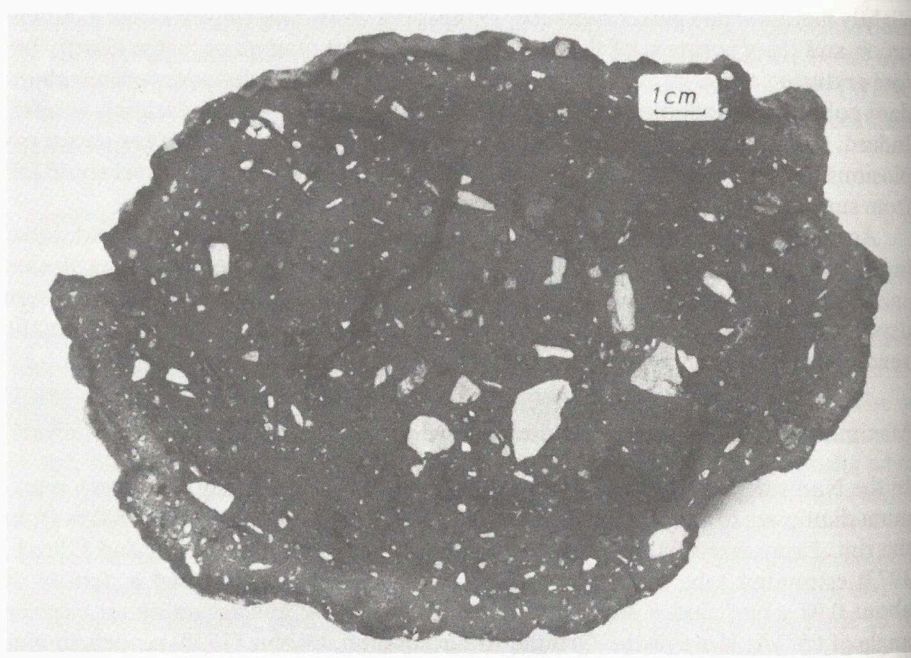
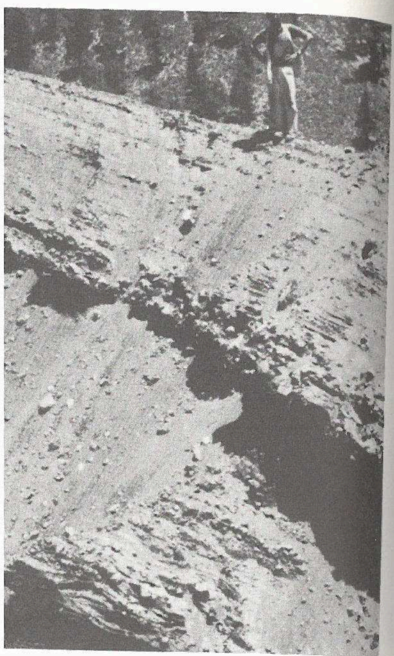


Fig. 9-7. Bomb from Quaternary phreatomagmatic deposits (Eifel) showing abundance of Devonian siltstone inclusions that have not been melted or strongly metamorphosed



ɔn of Duraznero Crater (La Palma, Canary
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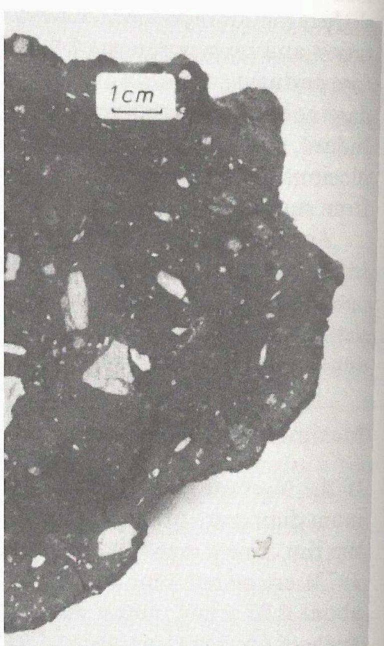

(Eifel) showing abundance of Devonian
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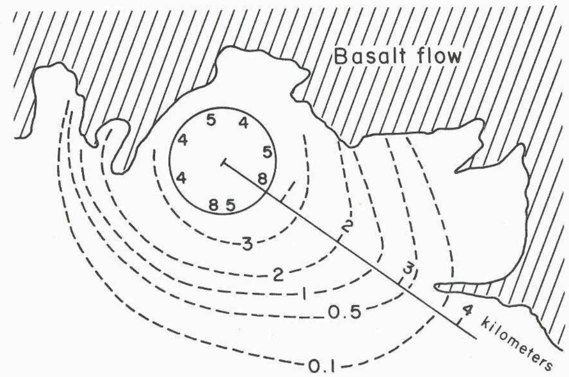
Fig. 9-8. Contours of maximum fragment sizes (in m). Hole-in-the-Ground (Oregon). (After Lorenz, 1971)



sures in the vent were over 500 b, ejection velocities of the largest blocks were 90–120 m/s and the fluid density was 0.04 g/cm^3. Similar ejection velocities of largest blocks are reported from the 1977 eruption of Ukinrek Maars, Alaska (Self et al., 1980).

Ultramafic Xenoliths

Many maar and tuff ring deposits contain abundant ultramafic rock fragments of different kinds thought to be derived from the mantle. These xenoliths have been used as evidence that maars result from explosions of magma rich in volatiles, especially CO_2 exsolved at great depth (McGetchin and Ullrich, 1973; Ringwood, 1975). However, if this were so, several conditions would have to be fulfilled. For example, there should be a close relationship between maar deposits and ultramafic nodules – which has not yet been demonstrated. Lava flows choked with ultramafic nodules are not uncommon, but many eruption centers in monogenetic volcano fields have not emitted lava flows. Secondly, magmas of maar-forming eruptions encompass a wide compositional range, while ultramafic nodules are typically restricted to alkalic mafic magmas, including kimberlites. Thirdly, little is known about the mechanism by which peridotite or other xenoliths become torn off conduit walls and incorporated into ascending magma. Even though mafic alkalic magmas are now thought to be generated at high CO_2 contents and high CO_2/H_2O ratios (Wyllie, 1979), and even though CO_2 exsolves at much greater depths than H_2O because of its higher partial pressure, no convincing cases have been made so far for fluidized magma particle-gas systems to have developed at depths exceeding a few kilometers.

In some cases, however, magma composition and maar-forming eruptions appear to be related. If so, it might be due to the rate of ascent of magma. Ascent is probably much faster for low viscosity, highly alkalic silica-undersaturated magma than for subalkalic tholeiitic magmas. Fast ascent would favor both nodule transport and, perhaps, the probability of explosive magma-water interactions (Schmincke, 1977a; Delaney, 1982). Kimberlite breccias which have been postulated to form by phreatomagmatic processes (Lorenz, 1975, 1980; Dawson, 1980), and which are known for their abundance of ultramafic nodules, may thus constitute an extreme end member in which very fast ascent rates favor both high transport capacities for nodules as well as high probability for magma-groundwater interaction in the upper crust. Only minor vesiculation of the magma would increase its surface area available for enhanced phreatomagmatic interactions.

Structures

Deposits of hydroclastic origin are characterized by well-developed beds ranging in thickness from a few millimeters to several tens of centimeters; most are less than about 10 cm thick. The abundance of thin beds presumably results from the large number of short eruptive pulses characteristic of hydroclastic eruptions (Figs. 9-9, 9-10). Layers vary from plane parallel beds to cross-bedded, lenticular beds that show scouring features, giving the misleading impression that they are reworked. However, transport directions radially outward from the crater are shown unambiguously by imbrication of platy fragments, cross-bedding geometry and isopachs. Cross bedding is discussed in the section on base surge deposits.

Penecontemporaneous Soft Sediment Deformation

Soft sediment deformation structures have been reported from hydroclastic deposits by several workers (e.g. Fisher and Waters, 1970; Lorenz, 1970, 1974; Schmincke, 1970; Heiken, 1971; Crowe and Fisher, 1973; Schmincke et al., 1973). The most common type resembles "convolute lamination" (Potter and Pettijohn, 1963), consisting of folded beds sandwiched between undeformed layers; deformed layers are several centimeters thick and may extend laterally for several meters. Two main explanations for the development of convolute laminations are: (1) gravity sliding of sloping water-saturated tephra (Heiken, 1971) and (2) shear-deformation caused by an overriding base surge flow (Fisher and Waters, 1970; Schmincke, 1970). A special type of convolute structure is the asymmetric "gravity" or "shear ripples", with wave lengths of 5–10 cm and amplitudes of 1-3 cm, that are inclined downslope in beds with initial dips of 5°–20° (Lorenz, 1974). Spectacular decollement folds (Fig. 9-11) have been observed 5 km from the source of tephra (Laacher See Volcano, Germany); downward sliding of tephra deposited on the slope of an older cone apparently produced the folds.

Vesicles (Gas Bubbles)

Vesicles, common in hydroclastic tuff beds of maar volcanoes (Lorenz, 1974), occur as subspherical voids, generally less than 1 mm but rarely exceeding 1 cm in diameter (Fig. 9-12). Most have smooth outlines and are coated by very fine-grained ash. Large vesicles are more irregular in shape than small vesicles and may consist of several coalesced bubbles. Vesicles are most common in beds showing soft sediment deformation but also occur in lahars, in tuff beds with mud cracks, and even in tuff plastered on vertical surfaces. Many tuff beds with vesicles contain accretionary lapilli (Self et al., 1980) which themselves may contain vesicles in their outer fine-grained layer (Lorenz, 1974, Fig. 7) or in their center. Vesicular tuffs

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0 cm and amplitudes of 1-3 cm,
s of 5°–20° (Lorenz, 1974). Spec-
bserved 5 km from the source of
rd sliding of tephra deposited on
folds.

ar volcanoes (Lorenz, 1974), oc-
ım but rarely exceeding 1 cm in
s and are coated by very fine-
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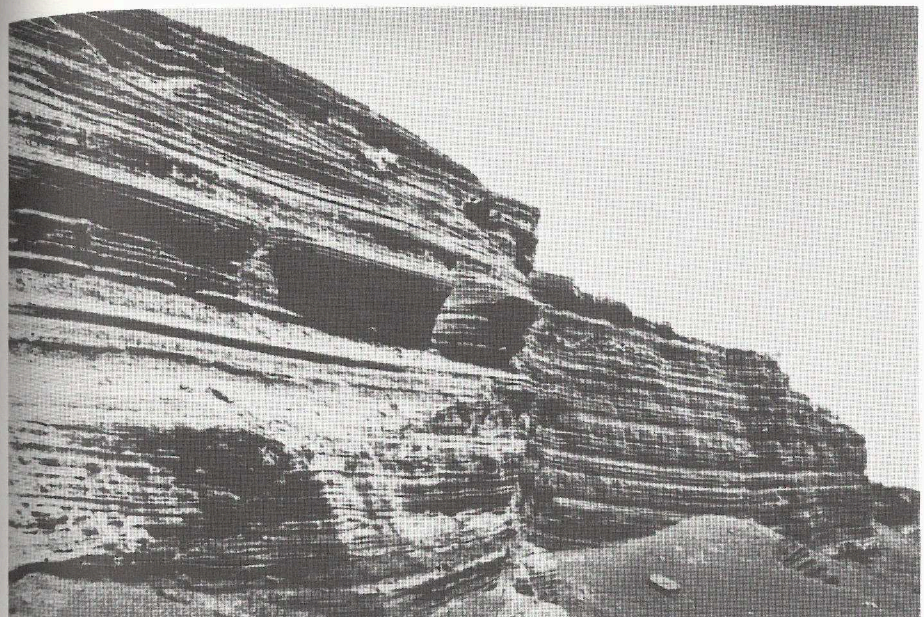


Fig. 9-9. Thin-bedded, horizontal base surge layers on rim of Salt Lake Crater (maar), Zuni Salt Lake (New Mexico) (Fisher and Waters, 1970)

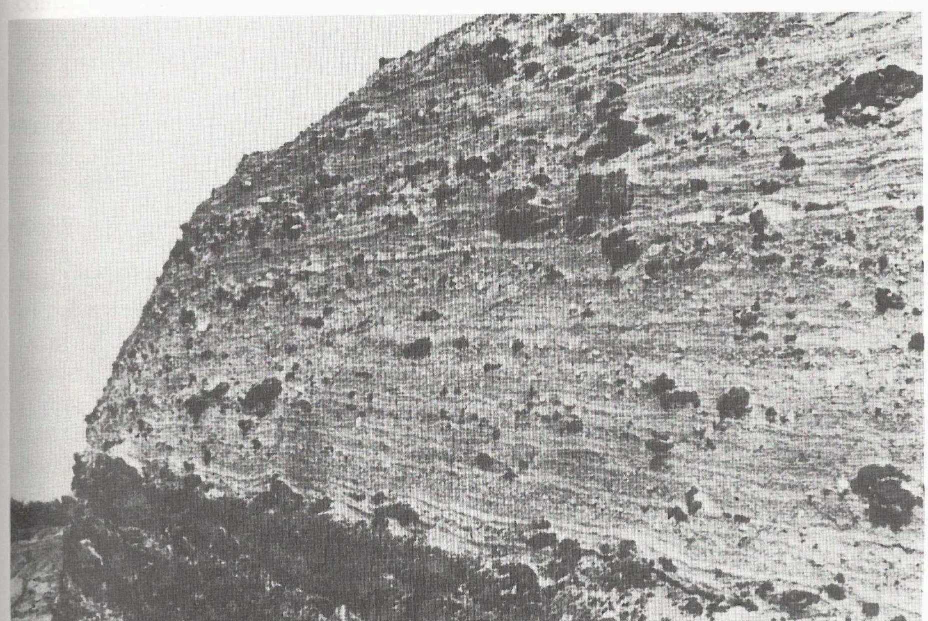


Fig. 9-10. Well-bedded lithic-rich tuffs formed by Vulcanian eruptions overlying slightly welded scoria deposits formed by Strombolian eruptions. Some bombs were erupted during the Vulcanian phase. Rothenberg volcano (Eifel, Germany). Large bombs 0.5 to 1 m in diameter

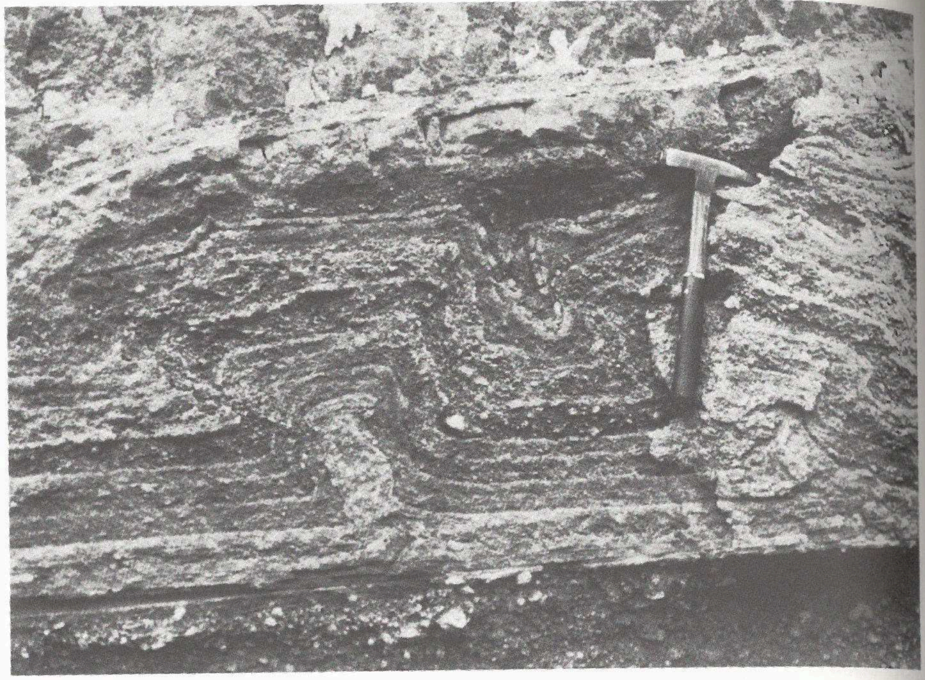


Fig. 9-11. Soft sediment deformation structure in base surge tuffs, many of which are vesiculated. Late Quaternary upper Laacher See tephra (Eifel, Germany)

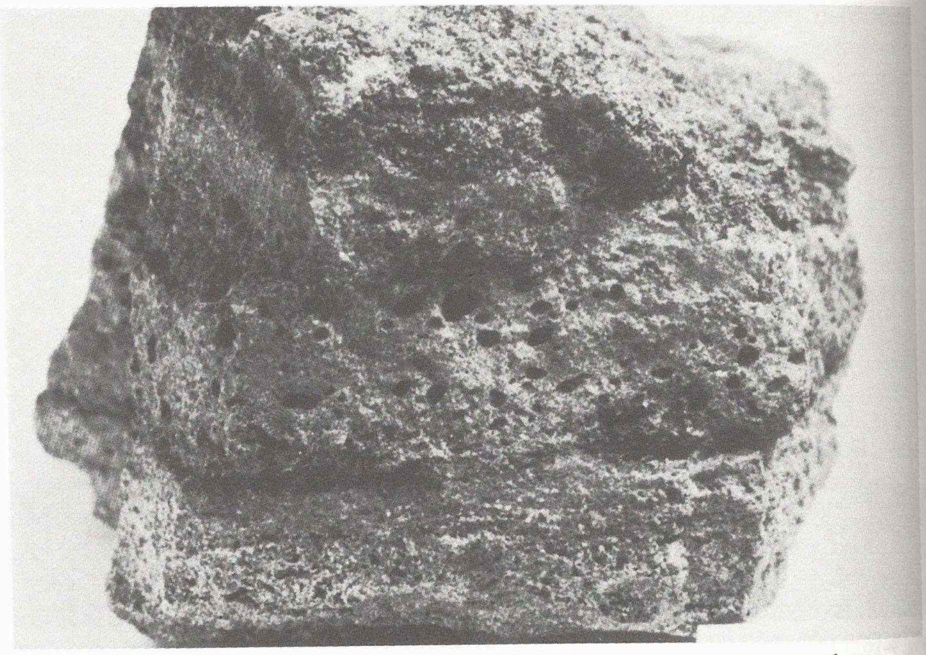
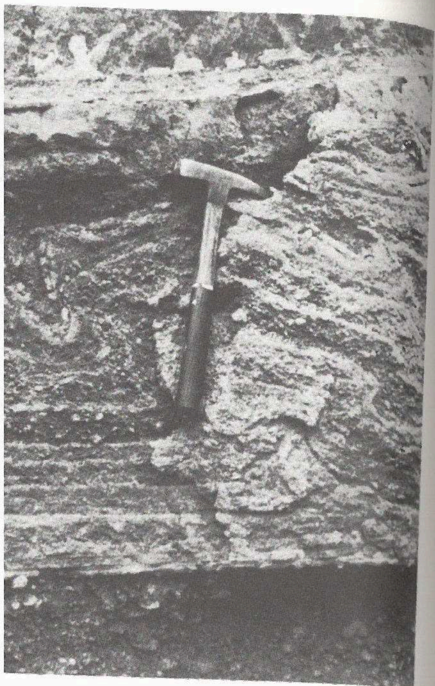
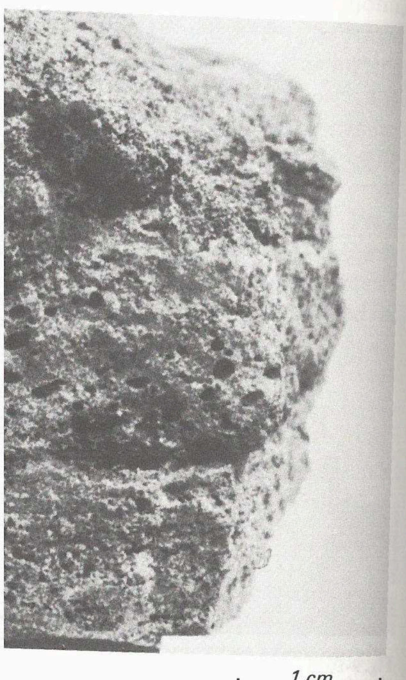


Fig. 9-12. Vesiculated tuff, base surge deposits, Vulcano (Eolian Islands) (Schmincke, 1977a)



rge tuffs, many of which are vesiculated. Late



olian Islands) (Schmincke, 1977a)

rarely occur more than 2–3 km from their source, although exceptionally they occur as far as 6 km (Laacher See, Germany). Vesicular tuffs are commonly more indurated than overlying and underlying vesicle-free beds.

Several sources of gas can account for vesicles in tuff: (1) gas within the fluidizing phase of the depositing system (derived from the eruptive center, from air incorporated during transit, or from rain falling during movement of the system), (2) gas given off by hot pyroclasts, (3) air rising from the underlying ground, or (4) water evaporating from snow or watersoaked soil beneath a layer of hot ash. Still another source may be (5) rain that turns to steam as it percolates downward into hot ash.

Several features prove that vesicle-bearing ash was water- or vapor-rich at the time of deposition: coating of vesicles with a film of clay or silt, association with soft sediment deformation structures, and preferential lithification of beds containing vesicles compared with associated nonvesicular beds. The amount of water necessary for soft sediment deformation in fine-grained sediment is about 15–20% (Heiken, 1971), far greater than saturation values for magmatic gases in basaltic magmas; thus the water or vapor phase must be mostly or entirely nonmagmatic. Moreover, vesicles can occur in tuff with mostly accidental clasts, thereby excluding them as a source of gas, although vesicles may form locally near large hot fragments exsolving gas or generating steam within a water-rich tuff matrix. Also, vesicles commonly occur several centimeters above the base of a bed, suggesting that air or gas did not rise from the ground below. Similar vesicles also may form in mudflow deposits (Sharp and Nobles, 1953; Bull, 1964).

Together with other criteria, the most common depositional mechanism of beds containing vesicles is probably a base surge resulting from phreatomagmatic eruptions (Lorenz, 1974; Self et al., 1980). However, at Augustine Volcano, Alaska vesicles also occur in fallout deposits. This fallout was probably in the form of "mud rain," a commonly reported event during volcanic eruptions (e.g. Macdonald, 1972). Tephra deposited from mud rains can develop vesicles from air or vapor trapped as bubbles within deposits of wet cohesive clasts.

Bedding Sags

Bedding sags, also known as "bomb sags" (Wentworth, 1926), form by the impact of ballistically ejected bombs, blocks, and lapilli upon beds capable of being plastically deformed (Fig. 9-13). They are characteristic of hydroclastic deposits and have been described from the deposits of many maar volcanoes, tiff rings, and tuff cones. Beds beneath the fragments may be completely penetrated, dragged down and thinned, folded, or show microfaulting (Heiken, 1971). Deformation is commonly asymmetrical, showing the angle and direction of impact if three-dimensional exposures are available.

Disruption of bedding also results when fragments fall into dry, noncohesive material. Dry material splashes out radially from the hole in rays or tongues or may construct a raised rim of ejecta around the depression with steep inward and gentle outward slopes (Hartmann, 1967). Because the particles do not stick together they are only affected by compressive forces. Unlike plastic beds, the noncohesive layers have no tensile strength. Thus, fragments beneath and in front of a projectile may be compressed together somewhat, but the force is rapidly dissipated by inter-

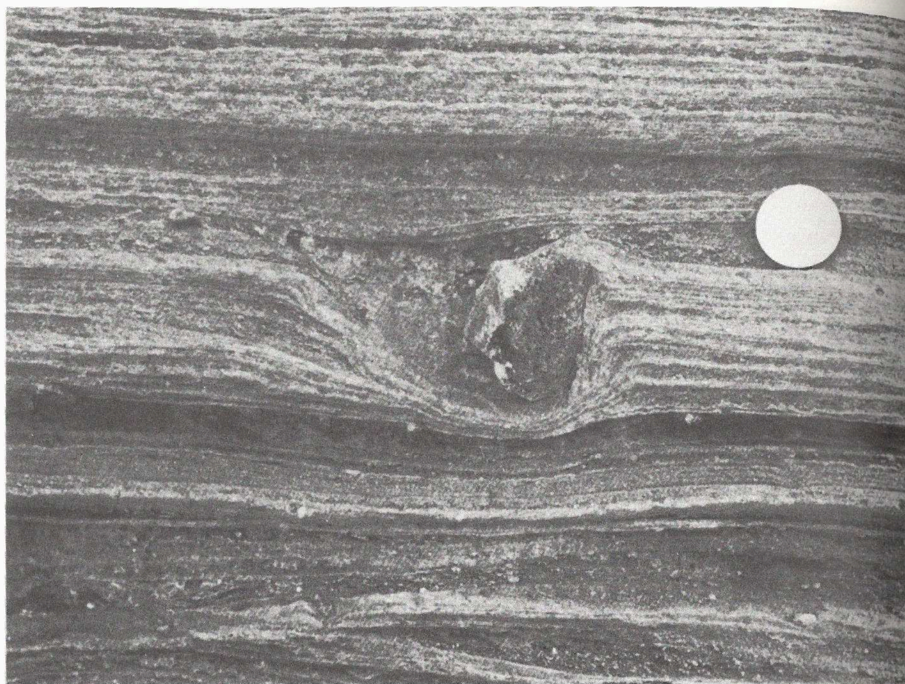


Fig. 9-13. Ballistically emplaced xenolith (transport direction from left to right) in thinly bedded base surge deposits, Duraznero crater (La Palma, Canary Islands)

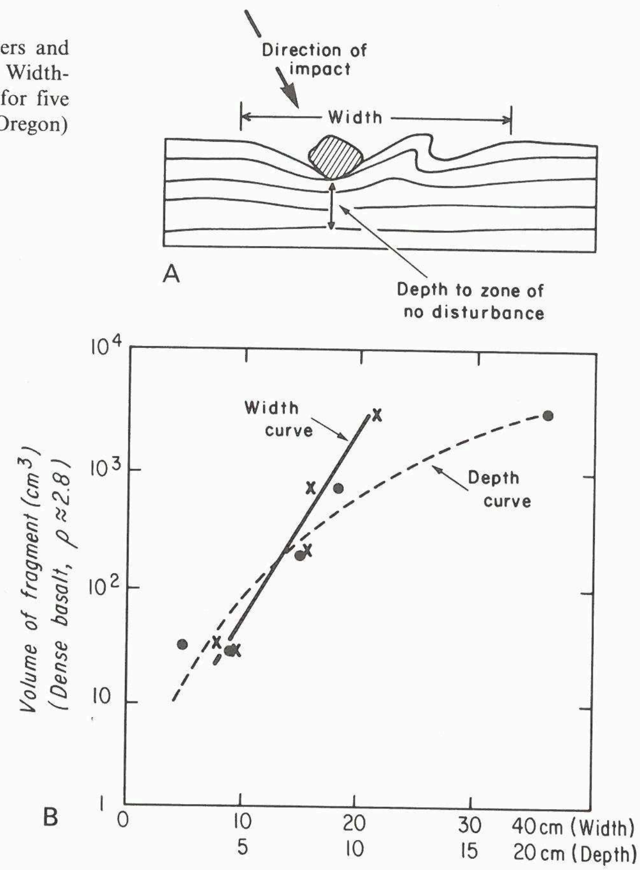
granular movement; beds do not deform plastically into folds or become stretched by flow.

The width and depth of disturbance due to an impact is in part a function of the momentum of the projectile, plasticity of the sediments and angle of impact (Fig. 9-14). Compaction of tuff also occurs as some water is forced out during impact and is more slowly displaced by the weight of the block. A few preliminary (unpublished) measurements on bedding sags at Prineville tuff cone in eastern Oregon suggest that the amount of deformation is a direct function of fragment mass. Width-depth ratios of pyroclastic bedding sags produced subaerially should greatly differ from ratios produced by dropstones in water, but such studies have not been published. Knowledge of the ratios, however, should aid interpretations of depositional environments.

Mudcracks

Penecontemporaneous mudcracks are observed in places on the surfaces of fine-grained hydroclastic deposits. This feature is reported by Lorenz (1974) at the Hverfjall tuff ring (Iceland) in two vesiculated tuff beds. We have also observed them in tuff cone deposits at Cerro Colorado, New Mexico, Koko Crater, Hawaii and Marteles Caldera, Gran Canaria.

Fig. 9-14. A Deformation parameters and terminology for bedding sags. **B** Width-depth data and fragment volume for five bedding sags, Prineville Tuff Cone (Oregon)



Base Surge Deposits

Base surges are a type of pyroclastic surge (Chap. 8) that form at the base of eruption columns and travel outward during some hydroclastic eruptions (Fig. 9-15). The name was originally applied to a surge which developed during the 1947 underwater nuclear test at Bikini Atoll (South Pacific) (Brinkley et al., 1950, p. 103–106), and subsequently was recognized in hydroclastic eruptions (Moore, 1967) following the 1965 phreatomagmatic eruption of Taal Volcano (Philippines) (Moore et al., 1966; Nakamura, 1966). Such flows appear to develop mainly by the collapse of vertical eruption columns as detailed in Waters and Fisher (1971). Condensed steam, an integral part of volcanic base surges, becomes thoroughly mixed with particles during flow. Water is trapped by surface tension as thin films around grains causing newly deposited material to be cohesive and behave plastically when deformed.

Single-stage fallback of eruption columns may occur in many cases, but the processes may be more complicated in others. During the 1975 eruption within Lake Ruapehu (New Zealand), for example, an initial base surge apparently developed from the pre-existing crater lake by spillout of water jets and expanding steam

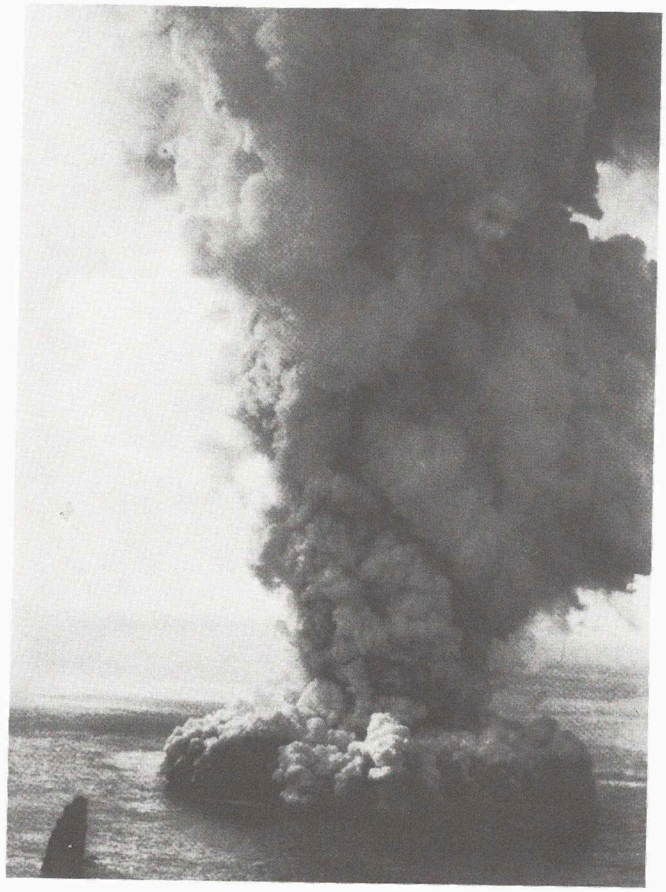
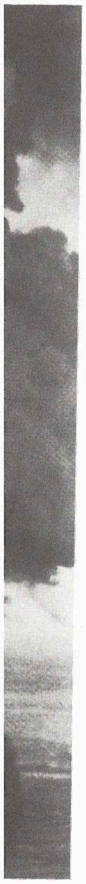


Fig. 9-15. Eruption column and base surge (diameter 270 m). Capelinhos (Azores). (Waters and Fisher, 1971)

(Nairn et al., 1979). Drainback of water into the lake was accompanied by collapse of the vertical eruption column to produce a "secondary" base surge composed of a dense aerosol of water droplets and debris. This surge moved at high enough velocities to surmount the rim of the crater 500 m above the lake and leave deposits on the outer slopes.

Eruptions that produce base surges involve release of large volumes of steam capable of supporting or fluidizing many of the particles in the surge.

Base surge deposits are poorly sorted and have an overall wedge-shaped geometry, decreasing logarithmically in thickness away from the source with local thickness controlled by topography (Wohletz and Sheridan, 1979). Distinct changes in facies and bed forms believed to be related to transport mechanism, load, and velocity of the surges are associated with decreasing thickness (distance). In places, the maximum radial distance attained by recognizable base surge deposits is about the same as the diameter of the crater (Table 9-3, Fig. 9-16), but at others, such as Laacher See, Germany (diameter = 2 km), base surge deposits occur 5 km or



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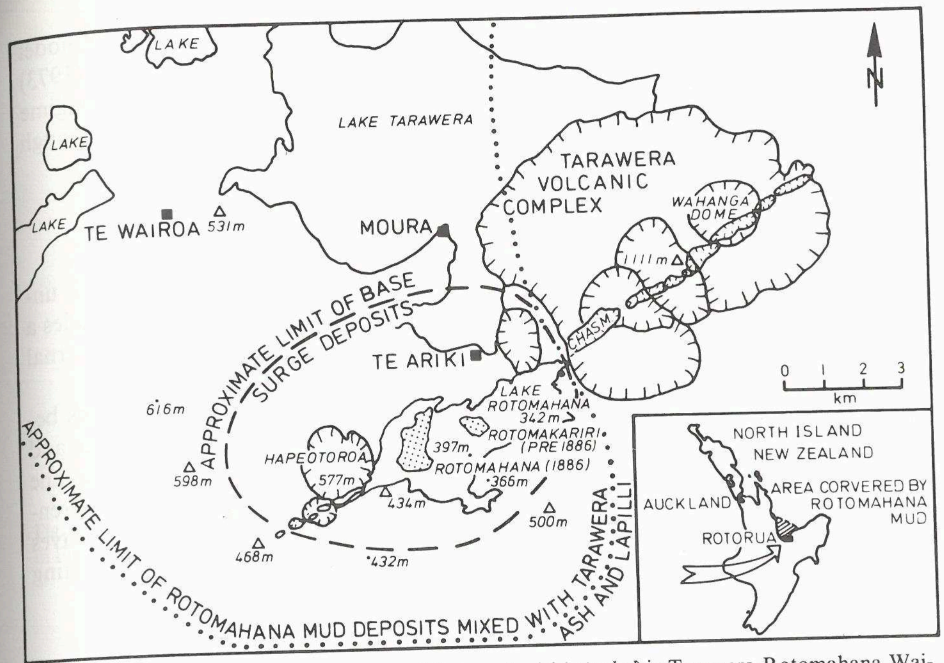

Fig. 9-16. Pre-eruption lakes Rotomahana and Rotomakariri (*stippled*) in Tarawera-Rotomahana-Waimangu area. Approximate limits of fallout tephra and surge deposits of Rotomahana are shown. (After Nairn, 1979)

Table 9-3. Crater size versus spread of surge deposit (Wohletz and Sheridan, 1979)

Deposit	Diameter of crater rim (km)	Maximum distance from crater rim (km)
Coronado Mesa, Arizona	0.5	1.1
Peridot Mesa, Arizona	0.55	1.25
Ubehebe Crater, California	0.80	0.82
Sugarloaf Mountain, Arizona	1.0	1.2
Elegante Crater, Mexico	1.6	> 1.1
Bishop Tuff, California	16 × 29	>15

more from source (Bogaard, 1978). Halemaumau, Hawaii, with a present-day diameter of about 1 km, was the site of phreatomagmatic eruptions in 1790 producing base surge deposits possibly as far as 10 km from the crater rim (Swanson and Christiansen, 1973).

Bed Forms from Base Surges

Bed forms occur as three main kinds – sandwave, massive and planar (plane parallel) beds (Schmincke et al., 1973; Sheridan and Updike, 1975), and are grouped into three facies types (Wohletz and Sheridan, 1979) related to a fluidization model of transport and deposition. Fisher and Waters (1970), Crowe and Fisher (1973) and Schmincke et al. (1973) have emphasized bed forms in terms of the flow regime concept (Chap. 5). These different approaches are treated separately although they are not mutually exclusive.

Sandwave Beds

The term sandwave bed (Sheridan and Updike, 1975) is applied to beds with undulating surfaces or surfaces inclined to the depositional substrate and includes a variety of bed forms such as surface dunes, antidunes and ripples, and internal cross laminations that make up dunes and ripples.

Sandwaves deposited by base surges have a wide range of characteristics believed to be related to the flow regime in which they were deposited (Fisher and Waters, 1970). Most workers believe that base surge bed forms develop within the upper flow regime, but lower flow regime forms might be present (Stuart and Brenner, 1979) (Fig. 9-17). As shown in Figure 9-18, these different types of sandwaves occur at Hunt's Hole (New Mexico) and are developed in a sequence suggesting an upward and lateral decrease in flow regime.

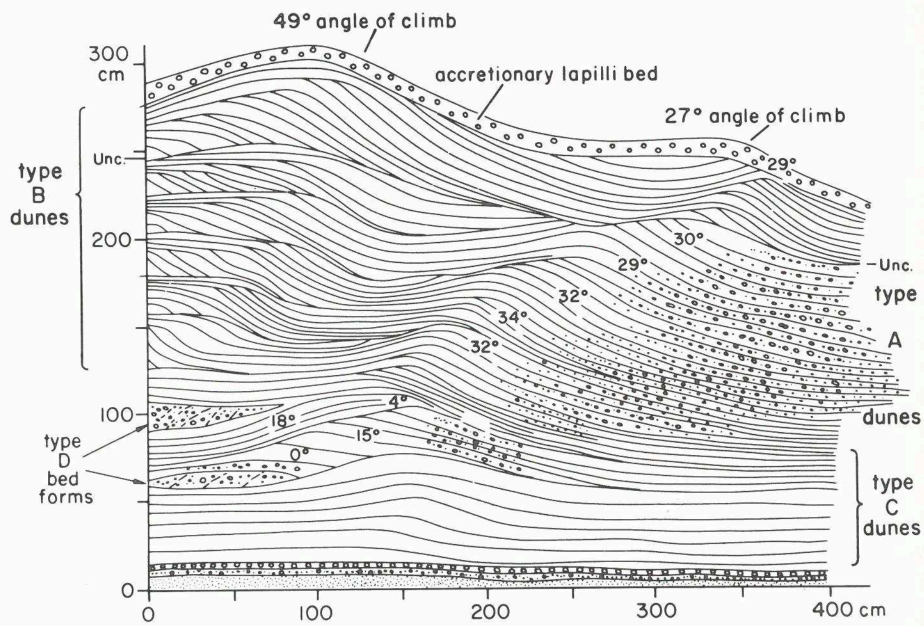

Fig. 9-17. Flow regime and bedforms in base surge deposits. *A* Lower-regime ripple drift dunes; foreset laminations approach angle of repose; backsets dip 15° or less toward source. *B* Lower-regime ripple drift dune forms with backset laminations horizontal to about 5° toward the source and foresets dipping up to 15°. *C* Upper regime antidunes. Backset laminations dip 5°–20° toward source, foresets are less than 15°. Laterally, *C* and *D* bedforms may be massive or occur as plane parallel beds. *D* Upper regime chute and pool bedforms. Backset (stoss) beds dip toward source at angles up to 55°; foreset (lee) strata are generally < 15°. (After Stuart, unpublished)

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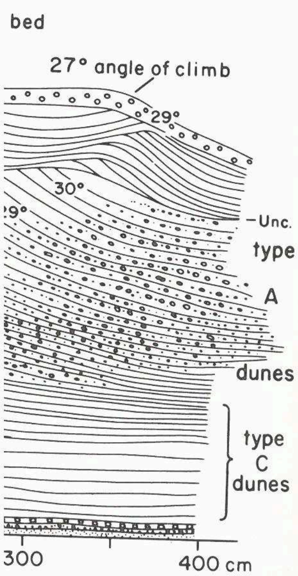

Lower-regime ripple drift dunes; foreset
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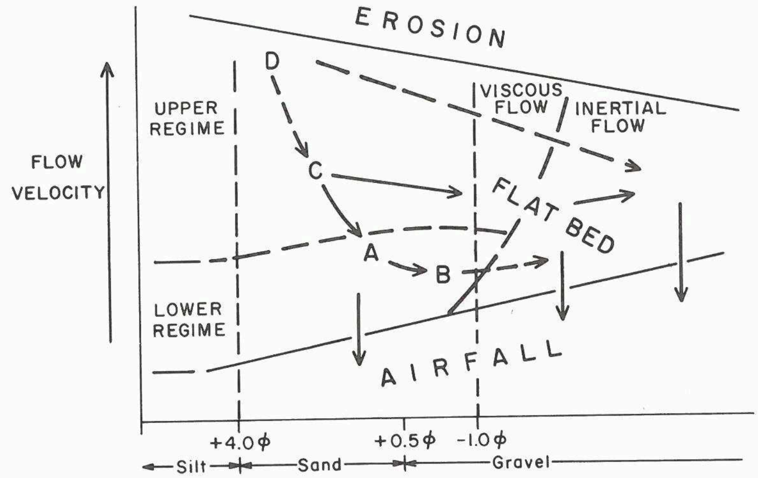

Fig. 9-18. Diagrammatic representation of bedform relationships. Bedforms *A*, *B* and *C* as explained in Fig. 9-17. (After Stuart, unpublished)

At other localities, lateral decreases in dune sizes and particle sizes and increases in sorting suggest decreasing velocities and probably flow regime, although lower-regime bed forms have heretofore gone unrecognized. At Taal Volcano (Philippines), base surge deposits within about 3 km of the vent are characterized by dunes oriented at right angles to movement directions of the base surges. The orientation of internal laminations show that the dunes migrated away from the explosion center. Wave lengths near the explosion center attained 19 m, systematically decreasing to about 4 m at 2.5 km from center. Northeast of the crater, the base surges were slowed by an uphill gradient, and wave lengths of dunes decreased rapidly (Moore, 1967). These relations suggest a direct relationship between wave length and velocity. The low dips of foreset (lee-side) laminations (10°–15°) indicate that the dunes did not advance by gravitational rolling of loose debris down advancing lee slopes, as is the case for desert sand dunes or low flow regime dunes formed in alluvial channels. Wave lengths of the Taal dunes vary directly with total thickness of the deposit, bedding thickness, size parameters of ash and the distance from the source (Table 9-4), suggesting that the carrying capacity of the Taal base surges decreased progressively with distance as velocity slowed.

Cross beds which occur in some flow units at Laacher See (Germany) also progressively change laterally from large dune-like structures characterized as chute and pool structures (Figs. 9-19, 9-20) (Schmincke et al., 1973) near the source to smaller more subdued antidunes farther from the source such as those at Ubehebe, California (Crowe and Fisher, 1973) (Fig. 9-21); the antidunes grade laterally into transitional low-amplitude structures that become plane-parallel beds about 5 km from Laacher See, and which appear to continue another 3–4 km from source. This progressive lateral change in bed forms, together with decreasing size and thickness parameters, are interpreted as reflecting a decrease in flow regime. Wave length probably depends on flow power, and wave height on grain size and volume of bed load. Thus, the downstream change suggests that the bed load was dropped rapidly

Table 9-4. Size parameters of samples from backset and foreset beds of dunes of base surge deposits. Taal volcano, Philippines (Waters and Fisher, 1971)

	Median diameter (Md_ϕ)	Sorting coefficient (σ_ϕ)	Distance from vent (km)	Approx. wavelength (m)	Approx. thickness of bedding set (m)	Approx. total thickness of deposits (m)
1. Backset	0.9	2.01				
Foreset	1.3	1.50	0.75	11.5	2.0	>4.0
2. Backset	2.4	1.65				
Foreset	2.9	1.24	1.0	9	1.0	~2.5
3. Backset	3.3	1.28				
Foreset	3.0	1.20	1.5	5.5	0.7	1.0
4. Backset	3.4	1.16				
Foreset	3.6	1.21	2.0	4	0.5	0.5

near the source and the energy of transport or capacity to carry a load then decreased more slowly.

Allen (1982) has criticized the hydrodynamic interpretation of sandwave bed forms in surge deposits as antidunes and chute and pool. In Allen's interpretation, these bedforms "record an unstable interaction between the moistened debris driven by the surge and a particle-capturing cohesive bed, that may have been independent of the Froude number" (Allen, 1982, p. 430). Allen subdivided bedforms and internal sedimentary structures in base surge deposits into (a) progressive bedforms – thought to be characteristic of relatively dry and/or hot flows, (b) stationary bedforms, and (c) regressive bedforms – with crests migrating upstream – thought to be deposited from relatively wet and cool flows. Previous authors have noted the problems of interpreting bedforms in systems characterized by cohesiveness. However, the correlation between the assumed wet regressive bedforms thought by Allen to be associated with accretionary lapilli and vesicle tuffs and the absence of these structures in progressive types is not supported by the evidence. At Laacher See, e.g., type C bedforms of Allen – the chute and pool structures of Schmincke et al. (1973) – occur only in the proximal base surge facies in coarse-grained relatively well-sorted deposits with very large wave lengths and amplitudes.

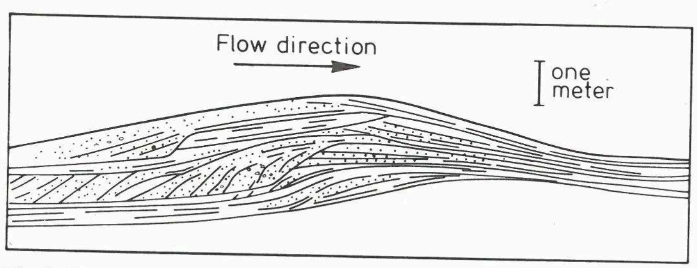

Fig. 9-19. Internal bedding of a chute-and-pool structure, Laacher See (Eifel, Germany). (After Schmincke et al., 1973)

foreset beds of dunes of base surge
l)

pprox. avelength 1)	Approx. thickness of bedding set (m)	Approx. total thickness of deposits (m)
.5	2.0	>4.0
'	1.0	~2.5
.5	0.7	1.0
	0.5	0.5

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Laacher See (Eifel, Germany). (After

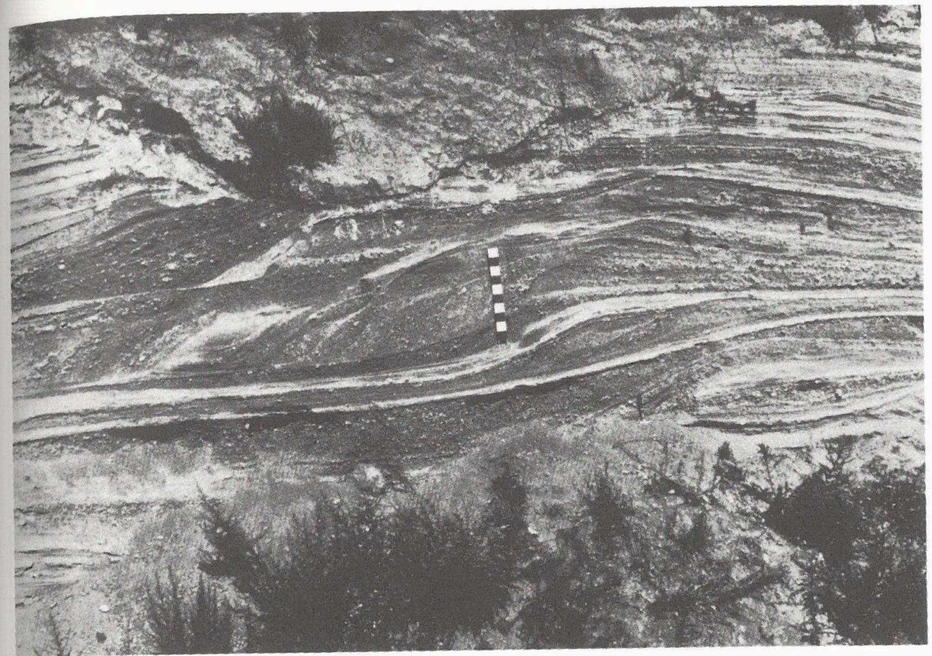

Fig. 9-20. Chute-and-pool structures in phonolitic base surge deposits representing a late eruptive phase of the Laacher See Volcano (Eifel, Germany). Scale in 10 cm intervals (Schmincke et al., 1973). Flow direction from *left to right*

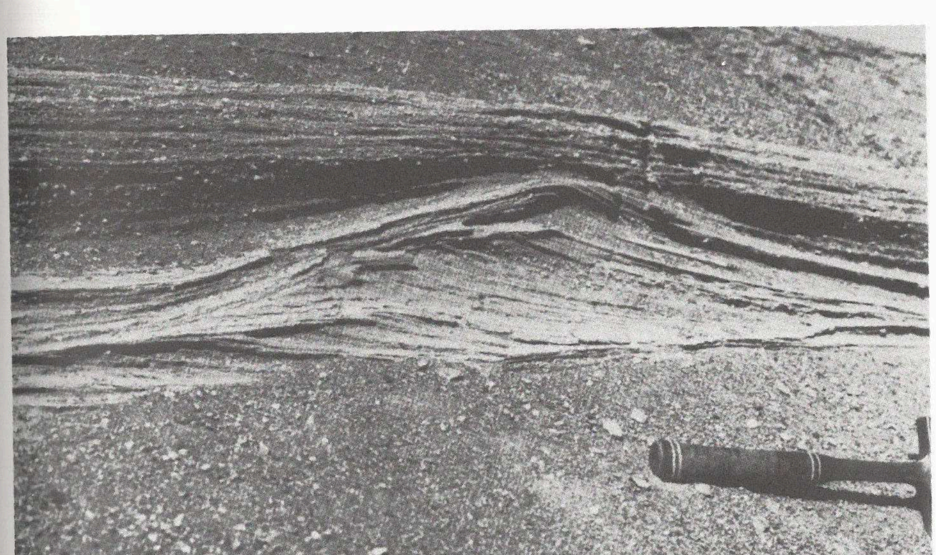

Fig. 9-21. Dune-like structure, Ubehebe Crater (California) showing build-up and migration from nearly flat beds. Flow direction from *left to right* (Fisher and Waters, 1970)

Downstream, in the more distal facies, the same beds develop progressive bedforms, the sediment being finer-grained and associated with accretionary lapilli and vesiculated ashes.

Although the hydrodynamic interpretation of the bedforms and internal structures of surge deposits is an open problem and incompletely studied as yet, we see no problem in supercritical flow being reached by surges.

Plane-Parallel Beds

Plane-parallel beds have upper and lower contacts which are generally planar and parallel to one another. Such beds in base surge deposits may be concordant with contiguous layers and normally or reversely graded, but unlike fallout layers, may erode into underlying beds. Sorting coefficients may be similar to fallout layers (Crowe and Fisher, 1973) but most surge deposits are more poorly sorted. In places, plane-parallel beds grade laterally from planar conformable sequences into zones of cross bedding, where they steepen into backset laminations (Schmincke et al., 1973). Plane-parallel beds tend to thicken within gentle lows and become thin and finer-grained over crestal parts of undulations, as do their internal laminae (Schmincke et al., 1973), rather than evenly mantling irregular underlying surfaces as is more common for fallout tephra. Platy fragments are imbricated or aligned roughly parallel to bedding surfaces. Internal laminae are commonly very subtly cross-bedded or lenticular over short distances (Fig. 9-22). Inversely graded plane-parallel beds suggest transport and deposition by flow, but are not unequivocal evidence inasmuch as some fallout beds are inversely graded (Chap. 6). Large blocks that rest on lower contacts without deformation are another indication of emplacement by flow, not fall.

Massive Beds

Massive beds usually are thicker, and more poorly sorted than plane-parallel beds or beds within sandwaves. They tend to be internally massive, but usually have pebble trains or vague internal textural variations giving a crude internal stratification that is either planar or wavelike (Fig. 9-23), and many massive beds have inversely graded basal zones. Sheridan and Updike (1975) and Wohletz and Sheridan (1979) postulate that massive beds are transported by a dense-phase fluidized surge and are transitional between sandwave and planar beds.

Bed Form Facies

The three facies defined by Wohletz and Sheridan (1979) are (1) sandwave facies (sandwave and massive beds), (2) massive facies (planar, massive and sandwave beds), and (3) planar facies (planar and massive beds). These facies systematically change laterally, with the sandwave facies dominating nearest the vent, massive facies at intermediate distances and the planar facies farthest from the vent (Fig. 9-24) at four volcanic sources described by them. Facies analyses of this kind may provide statistical summations of the dominant flow processes of many different flows through time at a particular locality, but we do not agree that they apply to the processes believed to occur laterally within a single flow.

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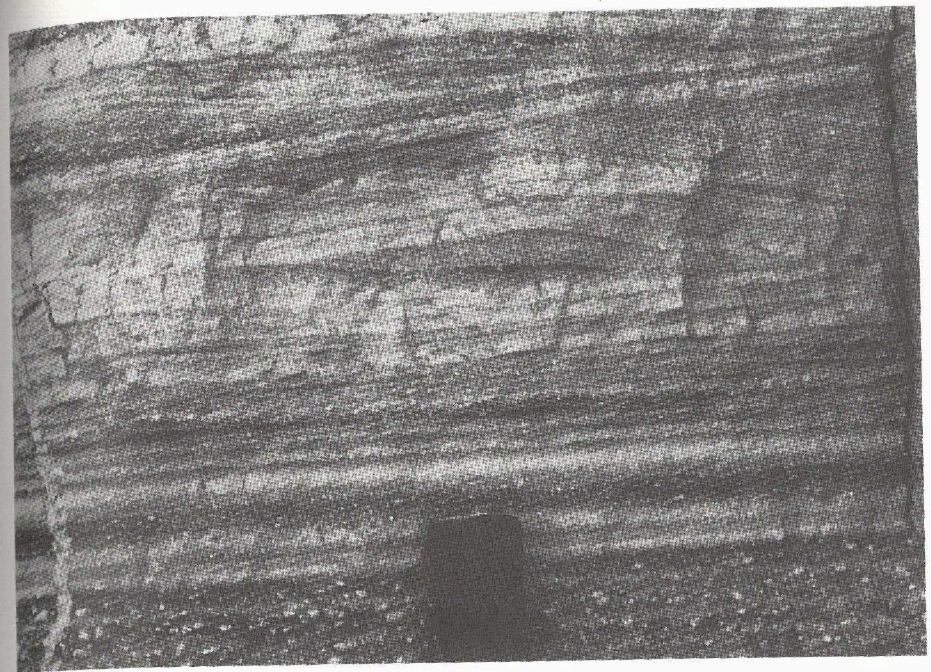

Fig. 9-22. Thin bedded base surge deposits, Ubehebe Crater (California), Flow direction from *left to right*

Fig. 9-23. Base surge bedforms at Laacher See (Eifel, Germany) (Schmincke et al., 1973). Flow direction from *right to left*

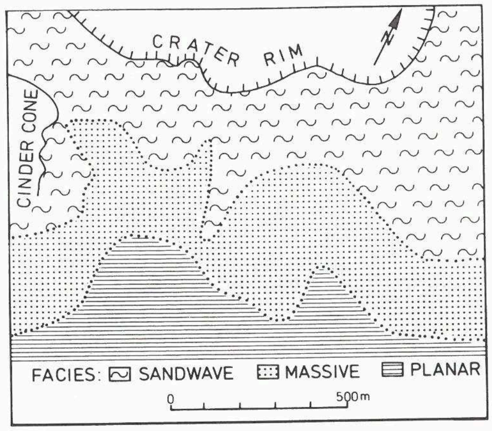

Fig. 9-24. Pyroclastic surge facies relationships at Crater Elegante (Sonora, Mexico). (After Wohletz and Sheridan, 1979)

Our alternative explanation for the three facies is (1) that massive and sandwave bed forms (i.e. sandwave facies) occur closest to the source because this is where the flows begin to separate by gravitational segregation into a laminar-flowing bedload and an overlying turbulent flow (Fisher et al., 1980; Fisher and Heiken, 1982); (2) planar beds from smaller flows begin to occur within the stratigraphic sequence at intermediate distances to give sequences with all three bed forms (i.e., the massive facies); and (3) massive beds within the distal planar facies might be thick planar beds. At Mukaiyama Volcano, Japan (Yokoyama and Tokunaga, 1978), plane-parallel to wavy beds (massive and sandwave bed forms?) occur closer to the source than large antidunes; farther from source the antidunes decrease in size. These relationships do not confirm or deny our alternative explanation for the development of the facies relationships shown by Wohletz and Sheridan (1979), but do point out that additional research is needed to resolve the many problems of facies.

U-Shaped Channels

U-shaped channels in base surge deposits, described by several authors (Losacco and Parea, 1969; Fisher and Waters, 1970; Mattson and Alvarez, 1973; Heiken, 1971; Schmincke, 1977b; Fisher, 1977; Nairn, 1979), are symmetrical in cross section, with curving bottoms that clearly cut underlying layers (Fig. 9-25). Most range from about 0.3 m to 7 m across and are a few centimeters to 3 m deep, but unusually large channels (30 m broad, 20 m deep) are reported by Losacco and Parea (1969), Mattson and Alvarez (1973) and Heiken (1971) and occur at Laacher See, Eifel, Germany (Fig. 9-25). The curving bottoms are best described as U-shaped, not parabolic or catenary curves, even though some are very broad in cross section. Infilling beds reflect the shape of the channels, but the curvature of individual beds decreases upward, and the final fill extends uniformly across the channel and is conformable with the sequence outside the channel. Thus, beds thicken toward the centers of channels and therefore do not resemble draped fall-out layers.

Fisher (1977) argues that the shape of the advancing head of a base surge and concentration of particles within the head is responsible for their U-shape. Rather

Fig. 9-24. Pyroclastic surge facies relationships at Crater Elegante (Sonora, Mexico). (After Wohletz and Sheridan, 1979)

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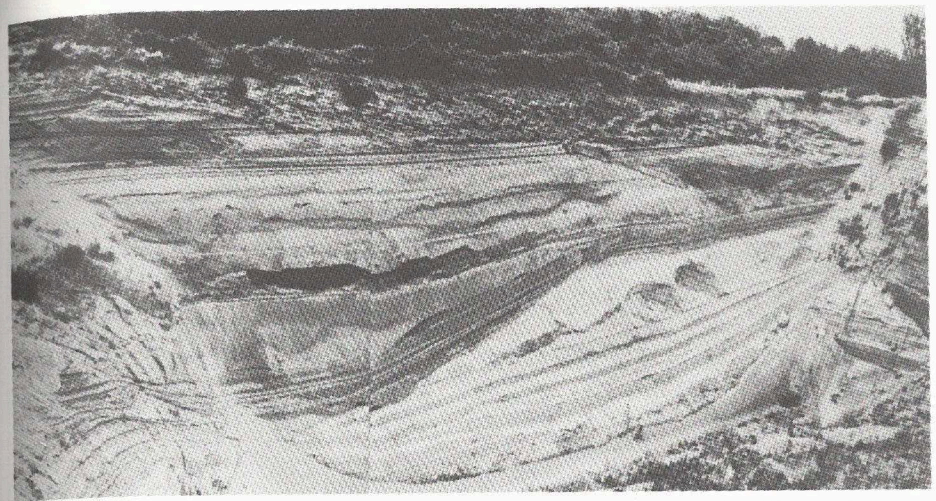

Fig. 9-25. Light-colored highly differentiated phonolitic fallout and ash flow tuffs with two minor unconformities formed by pyroclastic flows overlain by dark mafic phonolite Vulcanian base surge and pyroclastic flow deposits, the first base surges having eroded a major U-shaped channel. Laacher See (Eifel, Germany)

than having smooth, even fronts, base surges (as do nuées ardentes) develop secondary knuckle-like clefts and lobes that spread outward from the source, each lobe possibly being a separate complexly turbulent cell that joins the main body of the flow behind the advancing front. Moving down a widening slope of a volcano, individual lobes diverge to follow independent paths and carve diverging furrows straight down the slope. The concentration of particles within the turbulent cells is probably greatest along their central axes, where boundary effects are least and forward velocity is greatest. If pre-existing channels are present, the debris becomes more concentrated in the channels to increase the erosive capacity of the currents.

Maar Volcanoes

Maar volcanoes are low volcanic cones with bowl-shaped craters that are wide relative to rim height (Fig. 9-26). They were originally recognized as small subcircular crater lakes in the Quaternary volcanic district of the Eifel (Germany), the term being derived from the Latin "mare" for sea (Steininger, 1819). Classification, definition, and theories of maar origins are discussed by Noll (1967), Ollier (1967), Waters and Fisher (1970), Lorenz et al. (1971), Lorenz (1973; 1975), Pike (1974), and Wohletz (1980).

Classification

As modified from Lorenz (1973), we define the various kinds of maar volcanoes as follows:

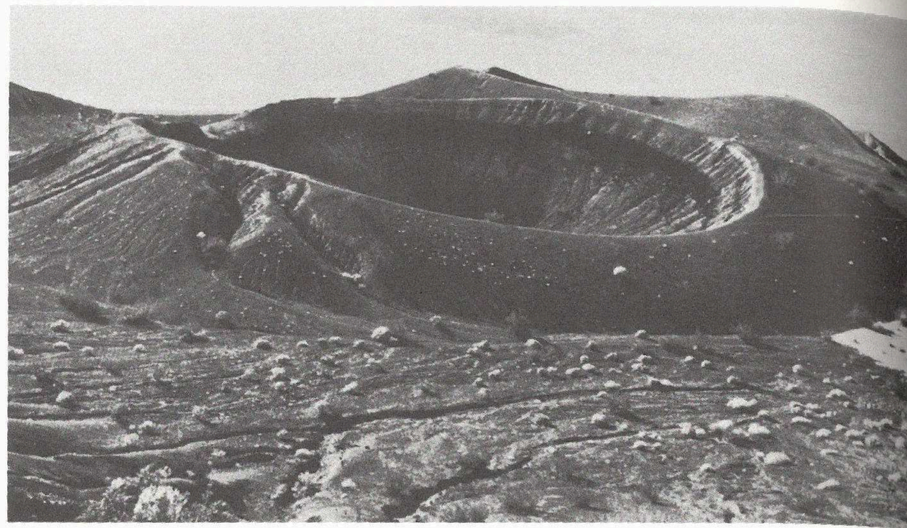

Fig. 9-26. Little Hebe Crater, tuff cone (Death Valley, California). Diameter about 100 m

Maar (sensu stricto): a volcanic crater cut into country rock *below* general ground level and possessing a low rim composed of coarse- to fine-grained tephra. They range from about 100 to 3000 m wide, about 10 to more than 500 m deep, and have a rim height of from a few meters to nearly 100 m above general ground level.

Tuff ring: a large volcanic crater at or *above* general ground level surrounded by a rim of pyroclastic debris (tuff or lapilli tuff), similar in diameter to maars. *Tuff cones* have higher rims, attaining heights of up to 300 m (Koko Crater, Hawaii), and are essentially tuff rings where volcanic activity was of longer duration. The distinction between tuff cones and tuff rings, however, becomes arbitrary where one side of a crater stands high and another side low. Aliamanu Crater, Hawaii, for example, would be classified as a tuff cone if viewed from the north and a tuff ring if viewed from the south, where it shares its rim with Salt Lake Crater, a low-standing tuff ring. Figure 9-27 depicts a hypothetical volcano showing the transition from a maar to a tuff cone based upon morphology.

Origin

Most maars result from hydroclastic eruptions (Lorenz, 1973; Kienle et al., 1980); wide craters develop from shallow explosions (Fisher and Waters, 1970), subsidence (Frechen, 1971; Noll, 1967) or a combination of both (Lorenz, 1973). Convincing evidence of a hydroclastic origin is that, in groups of nearly synchronous eruptive centers, those erupting on high ground form spatter or cinder cones, whereas associated eruption centers in valleys, depressions, on alluvial gravels or in coastal regions form maars, tuff rings or tuff cones (Lorenz, 1973). Juvenile clasts within their deposits are glassy, nonvesiculated, and have blocky shapes (Heiken, 1974), suggesting that magma was quenched prior to exsolution of vola-

ıia). Diameter about 100 m

into country rock *below* general
of coarse- to fine-grained tephra.
out 10 to more than 500 m deep,
early 100 m above general ground

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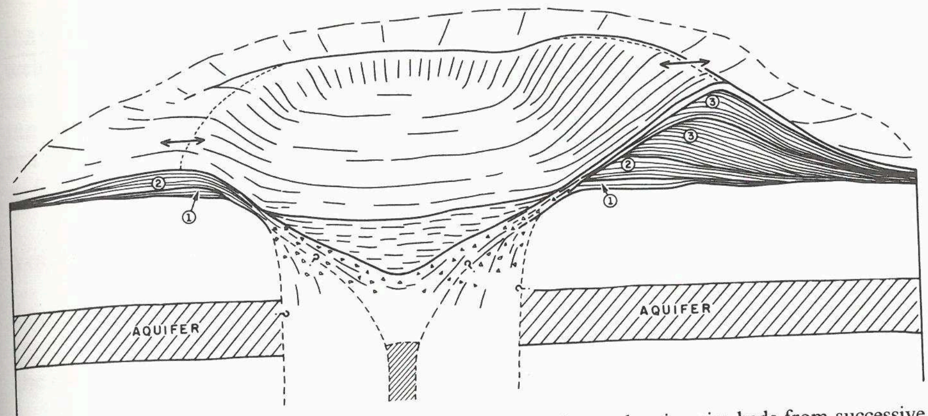


Fig. 9-27. Diagrammatic section of an asymmetrical maar volcano showing rim beds from successive stages of construction from explosion crater (*1*), tuff ring (*2*), and tuff cone (*3*). Anticlinal-form of rim beds is shown. Crater is filled with fall-back tephra and by lake and fluvial sediments. Conduit is shown as either the upper end of a diatreme formed by progressive collapse and partial blow-out of debris (Lorenz, 1973) or else is a funnel-shaped structure formed by explosion due to continuous mixing of magma and water at shallow depth. Unconformities on inner crater wall are from blast-erosion or collapse of steepened slopes

tiles, that breakage of glass resulted from thermal shock and (steam) explosions, and that the vapor and steam phase in the eruption column was partly or largely vapor from external water.

Wohletz and Sheridan (1983) conclude that tuff cones and tuff rings are distinct land forms that result from slightly different types of hydroclastic activity and they present a "hydroclastic continuum" of landforms from cinder cones to pillow lavas relating environments of eruption and mechanical energy of eruptions. According to them, tuff rings evolve through a stage of explosion breccia emplacement to a stage dominated by base surges which deposit thinly bedded layers. Tuff cones may be built when continuing activity evolves into a third stage, characterized by rocks emplaced by poorly inflated base surges and ballistic fallout. They relate these differences to water: melt ratios (Sheridan and Wohletz, 1983) based upon experiments with thermite-water systems (Wohletz, 1980). Fragmentation of melt attains maximum explosive energy when the water:melt ratio is about 0.5 for basaltic compositions. Initial ("vent-coring") eruptions with small ratios result in the formation of breccia with abundant cognate and accidental fragments. Increasing ratios cause development of expanded dilute surges which deposit thin-bedded layers, hence tuff rings. Still higher ratios produce "wetter" and denser eruption columns giving rise to poorly expanded surges, hence dominantly massive beds and tuff cones. The rates of magma and water influx controls the process, therefore such "cycles" may be interrupted, reversed or alternate. We have observed scoria blanketed by phreatomagmatic breccias from the same vent (Figs. 9-6, 9-10), but most commonly, tuff cones may have craters filled or partly filled with lava (Womer et al., 1980), agglutinated spatter and cinders (Prineville, Oregon). In some volcanic fields, many of the scoria cones contain deposits of phreatomagmatic origin commonly developed during their initial eruptive stages (Schmincke, 1977a).

Table 9-5. 20th century maar-forming eruptions (Kienle et al., 1980)

Maar	Maximum diameter (km)	Year	Source
Corral Quemado, Chile	≃1	1907	Illies (1959)
Stübel crater, Kamchatka[b]	1.5	1907	Vlodavetz and Piip (1959)
Falcon Is., Tonga Islands	1.5	1927	Hoffmeister et al. (1929)
Pematang Bata, Sumatra[c]	2.1, 1.0	1933	Stehn (1934)
Nilahue, Chile	0.8–1.4[a]	1955	Müller and Veyl (1957), Zuniga (1956), Illies (1959)
Iwo Jima[c,d]	0.035	1956	Corwin and Foster (1959)
Deception Island (5 maars)[c], Antarctica	0.3–0.5	1967	Schultz (1972)
Vlodavets, Radkevich (Tyatya volcano, Kuriles)[c]	0.3, 0.3	1973	Markhinin et al. (1974)
Taal	0.2	1965–66	Moore et al. (1966), Moore (1967)
Ukinrek (West and East Maars)[c]	0.17, 0.3	1977	Kienle et al. (1980)

[a] Varies with different author estimates
[b] This is proposed to be a maar by C. A. Wood (personal communication)
[c] Two or more craters formed during eruptive event
[d] One crater is proposed to be a maar; a second crater is a collapse pit

Traditionally, maars were thought to have originated by the explosive discharge of mantle-derived CO_2, as discussed in Chapter 4. However, even carbonatite maars, formed from magmas rich in CO_2, appear to occur only in lowland regions of the African Rift Valley where ground water is available, and are therefore of probable hydroclastic origin (Dawson, 1964a,b).

Further evidence for the central role of external water comes from observations of historic maar-forming eruptions (Table 9-5), especially the 1977 Ukinrek Maars, Alaska (Kienle et al., 1980; Self et al., 1980). Characteristically, maar-forming eruptions are accompanied by great volumes of steam and repeated short-interval blasts. Often maars occur in groups of two or more. Indeed some large tuff rings have scalloped shapes that may be caused by several closely spaced eruption centers and/or inward slumping of rims into repeated explosively evacuated central craters.

Dimensions

Areal Extent and Geometry

Compared to cinder scoria of similar volume, maar and tuff ring deposits usually extend farther from the eruption center. Scoria cones are built from vertical eruption columns composed mainly of juvenile bombs, lapilli, and ash that are deposited as spatter (agglutinate and agglomerate) and lapilli and ash layers rich in tachylite and scoria; fragments tend to follow ballistic paths, and the bulk of the material falls back near the vent. In maar volcanoes, much of the ejecta is finer-grained than in scoria cones and much may be transported by base surges. The depth of explosions is usually shallow, so that ejection angles are commonly lower than from scoria cones. The contrast between scoria cones and maar volcanoes is well seen when profiles of both are compared (Heiken, 1971). Abundant fines are carried far beyond the sites of eruption but are quickly dispersed and eroded. During the March 30, 1977 eruption of Ukinrek Maars, Alaska, for example, fine ash fell over 20,000–25,000 km^2 but significant ash accumulation was restricted to a radius of only about 3 km.

Volume

Maar craters are often thought to have larger volumes than the material ejected. However, volume estimates commonly have underestimated amounts eroded from the rim deposits and especially far-distant fallout deposits.

The main problem in determining the volume of magma represented by maar deposits is a realistic estimate of distant fallout material and of essential material hidden in the diatreme beneath the crater floor. Using a formula modified from Lorenz (1971), Mertes (1983) determined the volume of material ejected in three Eifel maars using the term

$$V_D = V_E(\varrho_E - \gamma\varrho_j) - V_C\varrho_B/\varrho_B - \varrho_D + \gamma\varrho_j$$

V_D = volume of essential material in vent
ϱ_E = density of ejecta
ϱ_j = density of essential material
ϱ_B = density of basement rocks
ϱ_D = density of vent filling
V_C = volume of crater in basement and
γ = amount of essential material given as percentage of total volume of ejecta V_E.

In this calculation, intrusives and larger blocks of country rock are neglected. The relationship

$$V_T = b \cdot R3,$$

where the coefficient b (varying between 0.109 and 0.074) decreases with increasing radius of maar (R), was used to estimate the total volume of ejecta, V_T, from all Eifel maars.

Maars and tuff rings encompass eruption centers with volumes up to 12×10^6 m^3 – the range for scoria rings and cones – as well as large centers with volumes between 15 and 30×10^6 m^3 (Mertes, 1983) (Fig. 9-28). Magma volumes may be related to composition. Large maars in the Eifel district, for example, are mostly composed of melilite nephelinite. The low viscosity and, at low pressure, the high volatile content of these magmas could retard freezing of the feeder dikes resulting in especially efficient discharge (Mertes, 1983). Volumes can be estimated from rim diameters as shown in Fig. 9-29.

Taking all factors into account, the total volume of ejecta from Ukinrek Maars is estimated at 10×10^6 m^3 (dense rock equivalent), substantially greater than the combined volume of the two fresh Ukinrek craters calculated at 4.3×10^6 m^3. Kienle et al. (1980) account for the excess ejecta volume (5.7×10^6 m^3) as juvenile fallout material – that part of the ejecta generally unaccounted for in pre-historic deposits of maars and tuff rings.

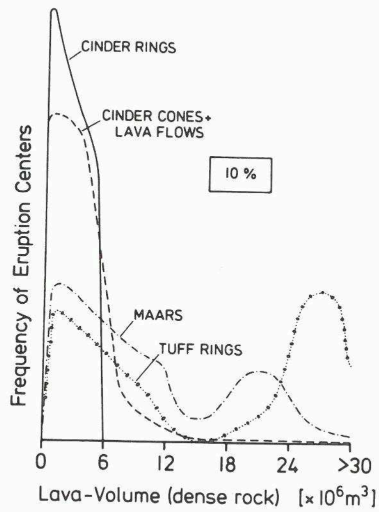


Fig. 9-28. Percentage of four major types of basaltic constructional volcanoes in Quaternary West Eifel volcanic field (about 240 eruptive centers) in relation to magma volume erupted (Mertes, 1983)

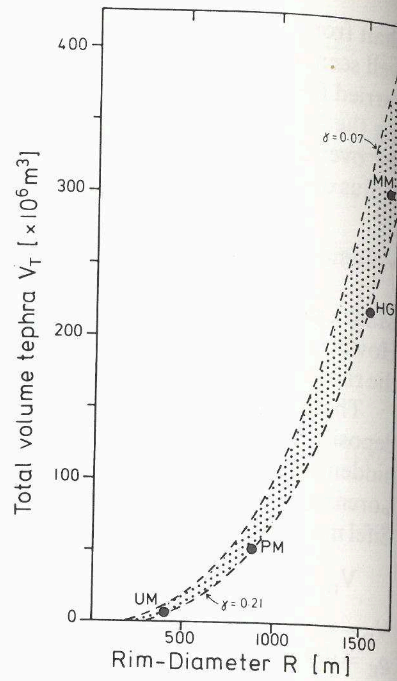
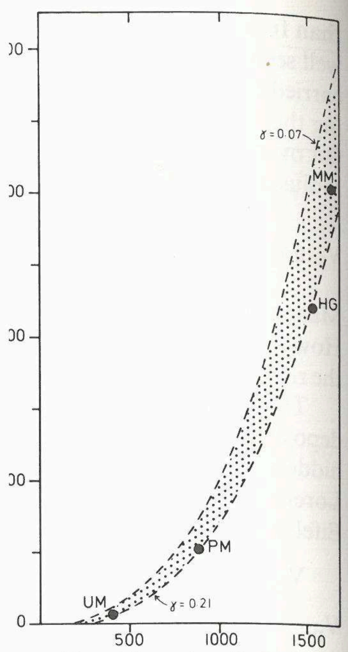


Fig. 9-29. Relationship between total tephra volume and rim diameter of maars. Bounding curves for Gamma = amount of essential clasts in fraction of total volume. *UM* Ulmener Maar, *PM* Pulvermaar, *HG* Hole-in-the-Ground, *MM* Meerfelder Maar (Mertes, 1983)

Chemical Composition

The composition of juvenile ejecta from maar volcanoes ranges widely, most being basaltic. In Iceland, for example, where about 20 maars were described by Noll (1967), most are tholeiitic, some with slightly alkalic basalt affinities. In central Oregon, USA, there are about 40 maars and tuff rings, mostly of high-alumina basalt (Heiken, 1971). In the Quaternary West Eifel volcanic field (Germany), most of the 70 maars, tuff rings, and tuff pipes are of melilite nephelinite and sodalite foidite composition, contrasting with the composition of the other 165 eruptive centers in the area (Mertes, 1983). The maars in East Africa are made up mostly of alkali basalt to nephelinite, some even of carbonatite (Dawson, 1964a, b). Maars of phonolitic (Schmincke et al., 1973; Schmincke, 1977a) and rhyolitic (Sheridan and Updike, 1975; Yokoyama and Tokunaga, 1978) composition have also been described.



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Littoral Cones

Littoral cones are mounds of hyaloclastic debris constructed by hydroclastic explosions at the point where lava enters the sea. Littoral cones belong to a group of craters that lack feeding vents connected to subsurface magma supplies (i.e., they are "rootless") and form where lava or pyroclastic flows move over small ponds of water, swamps, springs or streams as, for example, the pseudocraters in Iceland (e.g. Rittmann, 1962), and the phreatic explosion pits in pyroclastic flow deposits at Mount St. Helens (Rowley et al., 1981). Littoral cones commonly occur as crescent-shaped ridges (Fig. 9-30), breached by the source lava or more rarely as complete cones with craters occurring above lava tubes. Explosion centers are near or at the shore line, therefore about half of the radially exploded material falls into the sea, leaving a half-cone on land. A typical littoral cone is characterized by: (1) a wide crater (if the part missing at sea is reconstructed) and low rims; (2) steep inner slopes exposing truncated strata unconformably mantled by in-dipping strata; and (3) gentle outer slopes merging with the slope of the underlying terrain.

About 50 pre-historic littoral cones are known along the shores of Mauna Loa and Kilauea, Hawaii (Stearns and Macdonald, 1946). Twenty-one lava flows have entered the ocean along the shore of the Island of Hawaii between about 1800 and 1973 (Peterson, 1976, Table 1), but only four of the flows have developed littoral cones (Moore and Ault, 1965; Fisher, 1968a; Peterson, 1976).

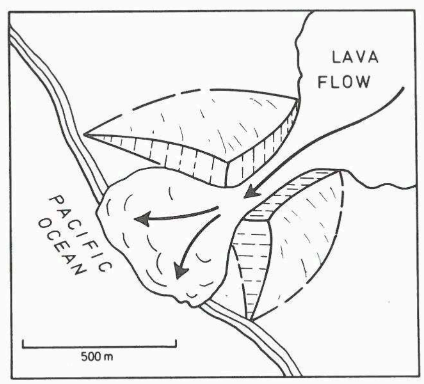

Fig. 9-30. Diagram of typical littoral cone

Deposits

Littoral cones are typically composed of hundreds of very poorly sorted, poorly defined beds ranging from a few centimeters to over 10 cm thick. They consist of fine- to coarse-grained ash, lapilli, and angular blocks up to 1.5 m and bombs to 1 m in longest dimension. Ash <0.062 mm ($>4_{\phi}$) diameter, however, is commonly no more than 5% of the total ash content. The ash is composed of sideromelane, tachylite, microcrystalline basalt and broken phenocrysts. Sideromelane fragments are predominantly broken fragments indicative of hydroclastic explosions (Heiken, 1974). Some layers contain accretionary lapilli and bedding sags, suggestive of abundant water vapor in the explosion clouds (Fisher, 1975).

Origin

The main conditions necessary for the construction of littoral cones appear to be a rapid delivery of large volumes of lava to the water (Moore and Ault, 1965; Fisher, 1968a) and confining conditions where water and lava can become repeatedly mixed. The abundance of beds within littoral cones and the height to which they can be built, about 100 m, suggests that conditions of confinement and water-lava mixing repeatedly occur. Lava tubes are absent at Puu Hou littoral cone, Hawaii, thus explosions probably occurred beneath confining lava crusts and rubble that continued to form as lava was fed to the ocean (Fisher, 1968a). Absence of lava tubes is characteristic of many pre-historic littoral cones on Hawaii's southeast coast, but some occur on top of lava tubes, a likely environment for confining conditions to occur. Indeed, the only observation of a littoral explosion not obscured by steam clouds (Peterson, 1976) indicates that lava tubes are important in the formation of some littoral cones.

Explosions that produce littoral cones may be caused by auto-catalysis (Fisher, 1975). The energy released by a given volume of water and lava during initial explosions may be great enough to cause further mixing and subsequent energy release of a somewhat larger volume of the two liquids in an exponential type of reaction (Colgate and Sigurgeirsson, 1973; see Chap. 4). Explosions subside as available lava is depleted by division into small droplets and expulsion from the mixing site, but take place again as lava continues to be delivered rapidly to the place of confinement under lava crust beneath water, or where water enters and comes in contact with molten lava within the confines of a lava tube at or below sea level.

Peperites

Peperites are a poorly defined group of volcaniclastic rocks that we include in the broad spectrum of hydroclastic rocks. The name peperite was used by Scrope (1862) for basaltic tuffs and breccias of the Limagne region of Central France. The rocks are composed of dark "basaltic" clasts in a light-colored marly to limy matrix – hence the name "pepper rocks." Scrope believed that they originated by explosions of Oligocene volcanoes, whose erupted material fell into lake water and mixed with sediment. Subsequently the peperites of Limagne were ascribed to explosive volcanic eruption or to the intrusion of basalt into wet sediment (Michel-Levy, 1890; Michel, 1953). Jones (1969b) suggested that peperites form by simultaneous sedimentation of epiclastic volcanic material together with lake sediments. The main argument against entirely epiclastic processes is the discrepancy between the fine-grained lake sediments and the coarse volcanic clasts. Moreover, the commonly glassy nature of the clasts is strong evidence for lava-water contact. The term peperite has been extended to other volcaniclastic breccias formed by mixing of hot lava and wet sediment during surface invasion unrelated to intrusive centers (Schmincke, 1967a). There are clear gradations from surface flows, through hyaloclastites, into peperites and finally "invasive flows" as lava flowed into sedimentary basins.