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Water and Magma Can Mix—
A History of the Concepts of Hydrovolcanism (1819-1980)

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INTRODUCTION AND SUMMARY

As geologists we rarely have completely original ideas, but continue to build on the work done by our predecessors. As geologic scholars we must look back to a legacy of observations and hypotheses as well as looking forward with new ideas and data sets. The evolution of ideas concerning hydrovolcanism (phreatomagmatic activity) has followed a well-beaten path from Darwin's observations in 1826 to the sophisticated observations, experimentation, and numerical modeling in the year 2000.

The first publications that describe maars (1819, in the Eifel) simply called them water-filled craters. During the voyage of the *Beagle* in the early 1830's, Charles Darwin not only determined that tuff rings were volcanoes but that they were formed along shorelines by the interaction of basaltic magmas with sea water. After those initial observations, few geologists questioned the process of magma/water interactions in eruptions that formed maars and tuff rings. A new variation on this theme came in 1885 when I.C. Russell studied tuff cones and rings in the Pleistocene lake basins of the Great Basin. From the late 19th Century to the present, the association of tuff rings with water has rarely been questioned.

In 1926 H. T. Stearns identified volcanoes on the Snake River Plain, Idaho as tuff rings, but the volcanoes had no obvious surface water source. With further work, Stearns identified aquifers below the volcanoes; he inferred that the aquifer provided water for magma/water interactions. The link between numerous hydrovolcanic volcanoes of the Roman province and large-scale aquifers below the coastal plain was made in the 1970's by Renato Funiciello and his colleagues at the University of Rome.

There have been disputes in some volcanic fields regarding the gas required for the energetic explosive eruptions that produced tuff rings and diatremes. For example, McGetchin and Ulrich, in 1973, proposed that ultramafic-xenolith-bearing tuff rings on the Colorado Plateau were produced during massive eruptions of CO_2. Lorenz and others have challenged this hypothesis, demonstrating a link to surface or ground water. It also has been proposed that both processes could have been responsible for some diatremes and maars. In any case, both proposed eruption phenomena have in common a large energy release.

Once the relationship between rising magma and water was established for the formation of tuff rings and maars, volcanologists began to concentrate on the physics of magma/water interactions and depositional processes that accompany hydrovolcanic eruptions. Major events that aided research on hydrovolcanic depositional mechanisms were the nuclear tests in the Pacific in the late 1940's, and the eruptions of Capelinhos Volcano, Azores (1958), Surtsey Volcano, Iceland (1963), and Taal Volcano, Philippines (1965). Observations of the nuclear tests and these eruptions led to research on volcanic density currents, variously called "base surge," "surge," or "ground surge." "Surges" have become an accepted process for deposition of many of the beds within tuff rings, whereas at one time most geologists thought that the main process was one of ballistic sedimentation.

More tuff ring studies, along with modeling and experiments, eventually led to closer analysis of eruption sequences in tuff rings. Of prime importance was the identification of "wet" and "dry" surge deposits and sequences where "drying out" accompanied the isolation of the vent from any water sources.

For those tuff rings (and even hydrovolcanic calderas) for which the water source was an aquifer or hydrothermal system, vertical variations in lithic clast populations through an eruption sequence allowed identification of the depth and nature of the aquifer. This approach, initiated in Italy by Funiciello, would eventually be used as a geothermal exploration tool in regions where hydrovolcanic activity sampled rock units below the volcano in aquifers that were potential hydrothermal systems.

An understanding of the genesis of maars and tuff rings has been an active topic for over 180 years. We learn by building on earlier studies. We too will eventually be part of the framework of understanding for hydrovolcanism, now bypassed by modern research on hydrovolcanism via numerical modeling and laboratory experiments—the subject of the next presentation.

OBSERVATIONS AND CONCEPTS

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 hydrovolcanism, which are defined as "explosions due to steam from any kind of water" (Wentworth, 1938). 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).

Stearns and Macdonald (1946) divided *hydroclastic* (hydro-explosion) eruptions into four categories:(1) *Phreatic* eruptions (explosions) are driven by the conversion of ground water to steam. Such steam explosions do not expel juvenile ejecta because ground water is vaporized only by heat or hot gases and not by direct contact with fresh magma. 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 or deep geothermal 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* or *hydrovolcanic* 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 (1977) 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* (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 as formed by the interaction of water and silicic magma Self and Sparks(1978). Such deposits differ from *Plinian* deposits principally by having 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 to 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.

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 (1961) and 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.

Deposits of Hydroclastic Eruptions

Many volcanic eruptions result from the interaction of magma and external water, but few volcanologists (e.g., Jaggar,1949) have emphasized the importance of non-magmatic water in volcanic eruptions. In our view, the general importance of external water in explosive eruptions is still underestimated. C. A. 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 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 such reservoir properties 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?

Components of Hydroclastic Deposits

Grain Size Distribution. The interaction of magma and water can produce extremely fine-grained clasts -- generally much finer than that produced by magmatic processes alone. Therefore, deposits from hydrovolcanic eruptions characteristically contain abundant fine-grained material, although coarse-grained lapilli- and tuff-breccias are common commonly interbedded with the fine-grained beds. 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 1mm in 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 Hydroclasts 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. Some vitric hydroclastic shards have abundant mosaic cracks indicating rapid chilling. 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). Honnorez (1966) and 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 caused by 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 of ash produced under water and typically occur in sea mounts 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, 1977). 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 (Schmincke, 1977; Lorenz, 1980). 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 Kramer (1970) first pointed out that surfaces of many lapilli and bombs from hydroclastic eruptions are characterized by what is described as a "cauliflower" texture (Lorenz, 1973)— a crackled or bread crust. This texture somewhat resembles that of bread crust bombs, but unlike breadcrust bombs formed by internal gas expansion, cauliflower bombs commonly have dense or only slightly vesicular interiors (Schmincke, 1977). 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, 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 reported by 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 other kinds of accretionary lapilli is the occurrence of vesicles in their outer layers (Lorenz, 1974) and in their core (Schmincke, 1977).

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 4mm to as much as 10cm 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.

Lithic (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. Accidental clasts also occur as inclusions in essential lapilli and bombs; most such clasts are only slightly metamorphosed (Schmincke, 1977) 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 (Stearns, 1950), 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 (200m deep; 600 x 1000m in diameter), maximum block diameters decrease from 3.3m at the rim to 70cm 2000m from the rim. From size relationships, Rohloff (1969; see also McGetchin and Ullrich, 1973) estimated that the eruptive fluid in Nanwaksjiak Maar had a density of about 0.01gcm3 and a surface velocity of about 500m/s assuming an ejection angle 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. Lorenz calculated that pressures in the vent were over 500 bars, ejection velocities of the largest blocks were 90-120m/s and the fluid density was 0.04g/cm3. 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. Outstanding examples of such deposits are Dreiser Weiher, Eifel, Germany (Stosch and Seck 1980; Aoki and Kushiro, 1968) and Salt Lake, Hawaii (White, 1966; Jackson and Wright, 1970; McGetchin and Ullrich, 1973). 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- sits and ultramafic nodules--which has not yet been demonstrated. Lava flows choked with ultramafic nodules are not uncommon, but many eruptive 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 alkaline 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 alkaline 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 related 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, 1977). 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 of Hydroclastic Deposits

Deposits of hydrovolcanic 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 hydrovolcanic eruptions 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.

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 (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 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.

Vesiculated Tuffs. 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. 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) or in their center. Vesicular tuffs 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 water- soaked 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 non-vesicular 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 can be mostly or entirely non-magmatic. 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, but such occurrences have been observed by us. 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, also known as "bomb sags" (Wentworth, 1926), form by the impact of ballistically-ejected bombs, blocks and lapilli into beds capable of being plastically deformed. They are characteristic of hydroclastic deposits and have been described from the deposits of many maar volcanoes, tuff rings and tuff cones. Beds beneath the fragments may be completely penetrated, dragged down and thinned, folded, or show micro- faulting (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, non-cohesive 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 non-cohesive 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 intergranular 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. 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.

Base Surge Deposits

Base surges are a type of pyroclastic surge that form at the base of eruption columns and travel outward during some hydrovolcanic eruptions. 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), and subsequently was recognized in hydrovolcanic 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 if 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 (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 500m 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-shape 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 but at others, such as Laacher See, Germany (diameter = 2 km), base surge deposits occur 5 km or more from source. 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 Left by 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), Fisher and Crowe (1973) and Schmincke et al. (1973) have emphasized bed forms in terms of the flow regime concept 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) The different sandwave types occur at Hunt's Hole (New Mexico) and were developed in a sequence suggesting an upward and lateral decrease in flow regime.

At other localities, lateral decreasing dune sizes and particle sizes and increasing 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 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 over short distances (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, 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 (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); 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 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). 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 though 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. 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 stuctures 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. Inversely graded plane parallel beds suggest transport and deposition by flow, but is not unequivocal evidence inasmuch as some fallout beds are inversely graded). 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, 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 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.

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 3 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 nor 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, 1977; Fisher, 1977; Nairn, 1979), are symmetrical in cross section, with curving bottoms that clearly cut underlying layers. Most range from about 0.3m to 7m across and are a few centimeters to 3 m deep, but unusually large channels (30 m across 20 m deep) are reported by Losacco and Parea (1969), Mattson and Alvarez (1973) and Heiken (1971) and occur at Laacher See, Eifel, Germany. The curving bottoms are best described as U-shaped, not parabolic 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 fallout 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 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. 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:

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 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 or where magma/water interactions were deep below the ground surface. 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.

Origin

Most maars result from “hydrovolcanic” or “phreatomagmatic” eruptions (Lorenz, 1973; Kienle et al., 1980); wide craters develop from shallow explosions (Fisher and Waters, 1970), subsidence (Frechen, 1962; 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 (Heiken, 1971; Lorenz, 1973). Juvenile clasts within their deposits are mostly glassy, nonvesiculated and have blocky shapes (Heiken, 1974), suggesting that magma was quenched prior to exsolution of volatiles, 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, 1981) 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, but most commonly, tuff cones with craters filled or partly filled with lava, 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, 1977).

Traditionally, maars were thought to have originated by the explosive discharge of mantle-derived CO_2, an interpretation advocated only recently (Barnes and McCoy, 1979). However, even carbonate maars, formed from magmas rich in CO_2, appear to occur only in low-land 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, 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 they 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.

Areal Extent and Geometry

Compared to cinder cones of similar volume, maar and tuff ring deposits usually extend farther from the eruption center. Cinder 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 cinder 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 cinder cones. The contrast between cinder 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 only to a radius of about 3 km.

Volume

In the past, maars were 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

VD=VE(QE-pQj)-VCQB/QB-QD+pQj

D=volume of essential material in vent
QE=density of ejecta
Qj=density of essential material
QB=density of basement rocks
QD=density of vent filling
VC=volume of crater in basement and
p=amount of essential material given as percentage of total volume of ejecta VE.

In this calculation, intrusives and larger blocks of country rock are neglected. The relationship $VT=b \times R^3$, where the coefficient b (varying between 0.109 and 0.074) decreases with increasing radius of maar (R), and was used to estimate the total volume of ejecta, VT, from all Eifel maars.

Maars and tuff rings encompass eruption centers with volumes up to $12 \times 10^6 m^3$ -- the range for cinder rings and cones -- as well as large centers with volumes between 15 and 30 $\times 10^6 m^3$ (Mertes and Schmincke, 1983. Magma volumes may be related to composition. Large maars in the Eifel district 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).

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 airfall material--that part of the ejecta generally unaccounted for in pre-historic deposits of maars and tuff rings.

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, U.S.A., 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, 1977) and rhyolitic (Sheridan and Updike, 1975; Yokoyama and Tokunaga, 1978) composition have also been described.

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 supples (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. Rittman, 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 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, 1968; Peterson, 1976).

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 1.5m and bombs to 1m in longest dimension. Ash <0.062 mm (>4 m) diameter, however, is commonly no more than 5% of the total ash content. The ash is composed of sideromelane, tachylite, microcrystalline basalt and broken phrenocrysts. 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, 1968) 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 100m, 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, 1968). 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 autocatalysis (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.

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.

REFERENCES

Allen, J. R. L., 1982. *Sedimentary Structures. Their character and physical basis.* Vol. II. Elsevier, Amsterdam, 663 pp.

Aoki, Kenichiro and Kushiro, Ikuo, 1968. Some clinopyroxenes from ultramafic inclusions in Dreiser Weiher, Eifel, Contributions to Mineralogy and Petrology, 18: 326-337.

Brinkley, S. R., Jr., Kirkwood, J. G., Lampson, C. W., Revelle, R., and Smith, S. B., 1950. Shock from undersater and underground blasts. In: The Effects of Atomic Weapons, Los Alamos Scientific Lab., Los Alamos, New Mexico. 83-113.

Bull, W. B., 1964. Alluvial fans and sear-surface subsidence in western Fresno County, California. U. S. Geol. Surv. Prof. Paper 437-4: 70 pp.

Carlisle, D., 1963. Pillow breccias and their aquagene tuffs, Quadra Island, British Columbia. J. Geol., 71: 48-71.

Colgate, S. A. and Sigurgeirsson, T., 1973. Dynamic mixing of water and lava. Nature, 244: 552-555.

Crowe, B. and Fisher, R. V., 1973. Sedimentary structures in base-surge deposits with special reference to cross bedding, Ubehebe Craters, Death Valley, California. Geol. Soc. Amer. Bull. 84: 663-682.

Darwin, C., 1891. Geological Observations. 3rd Ed., Smith, Elder and Co., London.

Dawson, J. B., 1980. *Kimberlites and Their Xenoliths.* Springer-Verlag, Berlin, 252 pp.

Dawson, J. B., 1964a. Carbonatitic volcanic ashes in northern Tanganyika. Bull. Volcanol., 27: 81-91.

Dawson, J. B., 1964b. Carbonatite tuff cones in northern Tanganyika. Geol. Mag., 101: 129-137.

Fisher, R. V., 1968. Puu Hou littoral cones. Geol. Rundsch. 57: 837-864.

Fisher, R. V., 1977. Erosion by volcanic base-surge density currents: U-shaped channels. Geol. Soc. Amer. Bull., 88: 1287-1297.

Fisher, R. V. and Waters, A. C., 1970. Base surge bed forms in maar volcanoes. Amer. J. Sci., 268: 157-180.

Fisher, R. V. and Heiken, G., 1982. Mt. Pelée, Martinique: May 8 and 20, 1902 pyroclastic flows and surges. J. Volc. Geotherm. Res., 13: 339-371.

Fisher, R. V., Smith, A. L., and Roobol, M. J., 1980. Destruction of St. Pierre, Martinique by ash cloud surges, May 8 and 20, 1902. Geology, 8: 472-476.

Frechen, J., 1971. Siebengebirge am Rhein, Laacher Vulkangebiet, Maargebiet der Westeifel, Vulkanologische-petrograaphische Exkursionen: 2nd Ed., Gebr. Borntraeger, Berline, 195 pp.

Funiciello, R., Locardi, E., Lombardi, G., and Parotto, M., 1976. The sedimentary ejecta from phreatomagmatic activity and their use for location of potential geothrmal areas. Intl. Cong. On Therm. Waters, Geotherm. Energy, Volcanol. Mediterranean Area, Athens.

Gary, M., McAfee, R., Jr., and Wolf, C. L. (eds), 1974. *Glossary of Geology*. Amer. Geol. Inst., Washington D. C., 805 pp.

Hartmann, W. K., 1967. Secondary volcanic impact craters at Kapoho, Hawaii and comparisons with lunar surfaces. Icarus, 7: 66-75.

Heiken, G. H., 1971. Tuff rings: examples from the Fort Rock-Christmas Lake Valley, south-central Oregon. J. Geophys. Res., 76, 5615-5626.

Heiken, G. H., 1972. Morphology and petrography of volcanic ashes. Geol. Soc. Bull., 83: 1961-1988.

Heiken, G. H., 1974. *An Atlas of Volcanic Ash.* Smithsonian Contr. Earth Sciences, 12: 101 pp.

Heiken, G. and Wohletz, K., 1985. *Volcanic Ash.* University of California Press, Berkeley, 246 pp.

Honnorez, J., 1961. Sur L'origine sur les hyaloclastites. Bull. Soc. Belge Geol., 70: 407-412.

Honnorez, J. and Kirst, P., 1975. Submarine basaltic volcanism: morphometric parameters for discriminating hyaloclastites from hyalotuffs. Bull. Volc., 39: 1-25.

Jackson, E. D. and Wright, T. L., 1970. Xenoliths in the Honolulu Series, Hawaii. J. Petrol., 11: 405-430.

Jaggar, T. A., 1949. Steam blast eruptions. Hawaiian Volcano Observatory, 4th Sp. Rpt, 137 pp.

Kienle, J., Kyle, P. R., Self, S., Motyka, R. J., and Lorenz, V., 1980. Ukinrek Maars, Alaska, I. April 1977 eruption sequence, petrology, and tectonic setting. J. Volcanol. Geotherm. Res., 7: 11-37.

Lorenz, V., 1970. Some aspects of the eruption mechanism of the Big Hole maar, Central Oregon. Geol. Soc. Amer. Bull., 81:1823-1830.

Lorenz, V., 1973. On the formation of maars. Bull. Volcanol., 37: 183-204.

Lorenz, V., 1974. Vesiculated tuffs and associated features. Sedimentology, 21: 273-291.

Lorenz, V., 1975. Formation of phreatomagmatic maar-diatreme volcanoes and its relevance to kimberlite diatremes. Phys. Chem. Earth, 9:17-27.

Lorenz, V., 1980. Explosive volcanism of alkali basaltic to kimberlithic melts. Lithos, 13: 217-219.

Lorenz, V., McBirney, A. R., and Williams, H., 1971. An investigation of volcanic depressions, Part III., Maars, tuff-rings, tuff-cones, and diatremes. NASA Progress Rpt. (NGR-38-003 012, Houston, Texas, 198 pp.

Losacco, U. and Parea, G. C., 1969. Saggio di un atlante di strutture sedimentaire e postsedimentarie osservate nelle piroclastiti del Lazio. Atti. Soc. Nat. e Matem. Di Modena, 94: 1-30.

Macdonald, G. A., 1972. *Volcanoes.* Prentice-Hall, Englewood Cliffs, N. J., 510 pp.

Mattson, P. H. and Alvarez, W., 1973. Base surge deposits in Pleistocene volcanic ash near Rome. Bull. Volc., 37: 553-572.

McGetchin, T. R. and Ullrich, G. W., 1973. Xenoliths in maars and diatremes with inferences for Moon, Mars and Venus. Jour. Geophys. Res. 78: 1832-1853.

Mertes, H., 1983. Aufbau und Genese des Westeifeler Vulkanfelds. Bochumer Geol. Und Geotechn. Arb., 9: 415 pp.

Moore, J. G., 1967. Base surge in recent volcanic eruptions. Bull. Volcanol., 30: 337-363.

Moore, J. G. and Ault, W. V., 1965. Historic littoral cones in Hawaii. Pacific Sci. 19: 3-11.

Moore, J. G. and Peck, D. L., 1962. Accretionary lapilli in volcanic rocks of the western continental United States. J. Geol., 70: 182-194.

Moore, J. G., Nakamura, K. and Alcarez, A., 1966. The 1965 eruption of Taal Volcano. Science, 151:955-960.

Muffler, L. J. P., White, D. E., and Truesdell, A. H., 1971. Hydrothermal explosion craters in Yellowstone National park. Geol. Soc. Amer. Bull., 82: 723-740.

Nairn, I. A., 1979. Rotomahana-Waimangu eruption, 1886. Base-surge and basaltic magma. N. Z. J. Geol. Geophys. 22: 363-378.

Nairn, I. A., Wood, C. P., and Hewson, C. A. Y., 1979. Phreatic eruptions of Ruapehu: April 1975. N. Z. J. Geol. Geophys. 22: 155-173.

Nakamura, K., 1966. The magmatophreatic eruptions of Taal Volcano in 1965, Philippines. Geology Magazine, 75: 93-104.

Nakamura, K. and Krämer, F., 1970. Basaltic ash flow deposits from a maar in West-Eifel, Germany. N. Jb. Geol. Pal. Mh., 8: 491-501.

Noll, H., 1967. *Maare und maar-ähnliche Explosionskrater in Island.* Köln Univ. Geol. Inst. Sonderveröffentlichungen, 117 pp.

Ollier, C. D., 1967. Maars, their characteristics, varieties, and definition. Bull. Volcanol., 31: 45-73.

Peterson, D. W., 1976. Processes of volcanic island growth, Kilauea Volcano, Hawaii, 1969-1973. In: Ferran, O. G. (ed) Proc. Andean and Antarctic Volcanology Problems, Intl. Assoc. Volcanol. Chem. Earth's Inter. Sp. Series, 172-189.

Pike, R. J., 1974. Craters on the Earth, Moon, and Mars: Multivariate classification and mode of origin. Earth Planet. Sci. Lett., 22: 245-255.

Potter, P. E. and Pettijohn, F. J., 1963. *Paleocurrents and Basin Analysis.* Springer, Berlin, 296 pp.

Ringwood, A. E., 1975. *Composition and Petrology of the Earth's Mantle.* McGraw-Hill, New York, 618 pp.

Rittmann, A., 1958. Il meccanismo di formazione dell lava a pillows e dei cosidetti tufi palagonitici. Atti. Acc. Gioenia 4: 310-317.

Rittmann, A., 1960. *Volcanoes and Their Activity.* John Wiley and Sons, New York, 305 pp.

Rohloff, K. J., 1969. Analysis of the exterior ballistics of block ejecta at Nanwaksjiak Crater, Nunivak Island Alaska. Air Force Inst. Tech., M. S. thesis, 118 pp.

Rowley, P. D., Kuntz, M. A., and MacLeod, N. S., 1981. Pyroclastic-flow deposits. In: Lipman, P. W. and Mullineaux, D. R. (eds), The 1980 eruptions of Mt. St. Helens , Washington. U. S. Geol. Surv. Prof. Paper 1250; 489-512.

Russell, I. C., 1885. Geological history of Lake Lahontan, a Quaternary lake of northwestern Nevada. Monograph - U. S. Geological Survey, M-0011, 288 pp,.

Schmincke, H.-U., 1970. Base surge—Ablagerungen des Laacher See-Vulcans. Aufschluss 21: 350-364.

Schmincke, H.-U., 1974. Pyroclastic Rocks. In Füchtbauer, H., *Sediments and Sedimentary rocks*. Schweizerbart'sch Verlagsbuchhandlung, Stuttgart, 160-189.

Schmincke, H.-U., 1977. Phreatomagmatische phasen in quartären vulkanen der Osteifel. Geo. Jahrb., 39:3-45.

Schmincke, H.-U., Fisher, R. V., and Waters, A. C., 1973. Antidune and chute and pool structures in the base surge deposits of the Laacher See area, Germany. Sedimentology, 20: 553-574.

Self, S. and Sparks, R. S. J., 1978. Characteristics of wide-spread pyroclastic deposits formed by the interaction of silicic magma and water. Bull. Volc., 41: 1-17.

Self, S., Kienle, J. and juot, J. P., 1980. Ukinrek Maars, Alaska, II. Deposits and formation of the 1977 craters. J. Volcanol. Geotherm. Res., 7: 39-65.

Sheridan, M. F., 1971. Particle-size characteristics of pyroclastic tuffs. J. Geophys. Res., 76: 5627-5634.

Sharp, R. P. and Nobles, L. H., 1953. Mudflows of 1941 at Wrightwood, southern California. Geol. Soc. Amer. Bull., 64: 547-560.

Sheridan, M. F. and Updike, R. G., 1975. Sugarloaf Mountain tephra—a Pleistocene rhyolitic deposit of base-surge origin. Geol. Soc. Amer. Bull., 86:571-581.

Sheridan, M. F. and Wohletz, K. H., 1981. Hydrovolcanic explosions I. The systematics of water-pyroclast equilibration. Science, 212: 1387-1389.

Stearns, H. T., 1926. Volcanism in the Mud Lake area, Idaho, American Journal of Science, 11, p. 353-363.

Stearns, H. T. and Macdonald, G. A., 1946. Geology and ground-water resources of the Island of Hawaii. Hawaii Div. Hydrogr. Bull. 9: 363 pp.

Stearns, H. T. and Clark, W. D., 1930. Geology and groundwater resources of the Kau District, Hawaii. U. S. Geol. Surv. Water Supp. Paper 616, 194 pp.

Stearns, H. T. and Vaksvik, K. N., 1935. Geology and ground-water resources of the Island of Oahu, Hawaii. Hawaii Terr. Dept. Public Lands Div. Hydrog. Bull. 1: 479 pp.

Steininger, J., 1819. *Geognistische Studien am Mittlrhein.* Mainz, 223 pp.

Stosch, H. G. and Seck, H. A., 1980. Geochemistry and mineralogy of two spinel peridotite suites from Dreiser Weiher, West Germany, Geochimica et Cosmochimica Acta, 44: 457-470.

Stuart, C. J. and Brenner, M. G., 1979. "Low-regime" base surge dunes—An example from Kilbourne and Hunt's Holes, south-central New Mexico (abs). Geol. Soc. Amer. Abs. With Programs, 11: 525.

Swanson, D. A. and Christiansen, R. L., 1973. Tragic base surge in 1790 at Kilauea Volcanoe. Geology, 1: 83-86.

Tazieff, H., 1972. About deep-sea volcanism. Geol. Rundsch., 61: 470-480.

Walker, G. P. L., 1969. The breaking of magma. Geol. Mag., 106: 166-173.

Walker, G. P. L., 1973. Explosive volcanic eruptions—a new classification scheme. Geol. Rundsch., 62: 431-446.

Walker, G. P. L. and Blake, D. H., 1966. Formation of a palagonite breccia mass beneath a valley glacier. Q. Jour. Geol. Soc. London, 122: 45-61.

Walker, G. P. L. and Croasdale, r., 1972. Characteristics of some basaltic pyroclastics. Bull. Volc., 35: 303-317.

Waters, A. C. and Fisher, R. V., 1971. Base surges and their deposits: Capelinhos and Taal volcanoes. J. Geophys. Res., 76: 5596-5614.

Wentworth, C. K., 1926. Pyroclastic geology of Oahu. Bishop Museum Bull. 30, 121 pp.

Wentworth, C. K., 1938. Ash formation of the Island of Hawaii. 3rd Sp. Rpt., Hawaiian Volcano Observatory, Honolulu, Hawaii, 183 pp.

White, R. W., 1966. Ultramafic inclusions in basaltic rocks from Hawaii, Contributions to Mineralogy and Petrology, 12: 245-314,

Wohletz, K. H., 1980. Explosive hydromagmatic volcanism. Arizona State Univ. Ph.D. diss., 303 pp.

Wohletz, K. H. and Sheridan, M. F., 1979. A model of pyroclastic surge. Geol. Soc. Amer. Spec. Paper 180: 177-194.

Wohletz, K. H. and Sheridan, M. F., 1983. Hydrovolcanic explosions II. Evolution of basaltic tuff rings and tuff cones. Amer. J. Sci., 283: 385-413.

Wyllie, P. J., 1979. Magmas and volatile components. Amer. Mineral., 64: 469-500.

Yokoyama, S. and Tokunaga, T., 1978. Base-surge deposits of Mukaiyama Volcano, Nii-jima, Izu Islands. Bull. Volcanol. Soc. Japan, 23: 249-262.

TABLE— CHRONOLOGY

Table—Water and Magma Can Mix—
A history of the concepts of Hydrovolcanism, (1819-1975)

Scientist	Year	Significant Concepts and Observations (1819-1975)
Descriptions and early interpretations	-------	---
J. Steininger	1819	Recognition as crater lakes
C. Daubeny	1826	Definition of "maar" as a lake, rather than a volcano.
Charles Darwin	1836	Galapagos. Descriptions of tuff rings; erupted immersed in the sea; eruptions of water-saturated tuff; alteration to "palagonite" (did all this in 5 days!)
Sartorius von Waltershausen	1837 & 1846	Sicily, Iceland; ashes from submarine eruptions; coined the terms 'palagonite' and "sideromelan'
James Dana	1849	Hawaii. Descriptions of tuff rings and cones on Oahu.
A. Humboldt	1865	Depressions formed by violent explosions
Charles Lyell	1872	Review of the subject of tuff cones and rings
Johnston-Lavis	1884	Concepts of hydrovolcanism at Vesuvius
Israel C. Russell	1885	Soda Lakes, Nevada;
Geike	1885	German maars.
Ordoñez	1906	Puebla, Mexico
Suess	1909	Link to diatremes
Darton	1916	Review of "explosion craters;" Zuni Salt Lake
Stearns	1926	Idaho; Snake River Plain. Inferences regarding the role of ground water in mixing with rising magma.
Martin Peacock	1926	Palagonite formation
Fenner	1921	Craters and lakes of Mt. Gambier, Australia
Firth	1930	Tuff rings around Auckland, NZ
Howell Williams/Alexander MacBirney	1936/ 1959	Hopi Buttes; diatreme/maar links; cratering "room" problem
Research Hiatus for WW		

II	--------	--
Cotton	1944	Geomorphology of maars; formed by sporadic explosions
Advances in the Understanding of the Physical Processes of Hydrovolcanism	-------	--
Jean Goguel	1952?	Physics of phreatic explosions
Gene Shoemaker	1957	Wind-blown ash to form dunes around tuff ring (Zuni Lake, NM)
Müller and Weyl	1957	Eruption of Nilahue, Chile
Corwin et al.	1957	Phreatic craters on Iwo Jima
Corwin and Foster	1956	Observed eruptions and crater formation, Iwo Jima
Frederico Machado	1962	Mechanisms; Capelinhos, Azores
Harold Stearns	1963	"Drying out" of tuff rings as water supply diminishes.
Norm Peterson and N. V. Groh	1963	Relation between magma/water interactions and basaltic volcanism across large lake basins, south-central Oregon
Sigurdur Thorarinsson	1964	Evolution of a tuff ring/scoria cone; Surtsey, Iceland.
Arthur Holmes	1965	Explosion vents and fluidization craters
Kristjan Saemundsson	1967	"Current bedding" in phreatomagmatic deposts that have reached or exceeded lake level.
Jim Moore,	1967	Surge bed deposition, Taal, Anak Krakatau; nuclear weapons analogy
Dick Hay and A. Iijima	1968	Palagonitization processes, Koko Craters, Hawaii
R. V. Fisher and Aaron Waters	1969 & 1970	Surge bed deposition, Taal, Capelinhos, Ubehebe Crater, Zuni Salt Lake, Salt Lake Craters; nuclear weapons analogy (Fisher was an observer)
E. H. Francis	1970	Subsidence in maar/tuff ring formation
Hans-Ulrich Schmincke et al.	1973	Surge processes, Laacher See
Grant Heiken	1971	Link of lines of scoria cones/tuff rings/tuff cones in a Pleistocene lake in Oregon; role of water—from dry ground into a lake with increasing depth. Surges, "wet" and "dry" beds.
Enrico Bonatti and Haroun Tazieff	1970	Hypothesis that some guyots are tuff cones
Volker Lorenz	1971	Ballistics, maar crater formation, role of ground water and calculation of kinetic energy. Links between diatremes and tuff rings in the Saar-Nahe
Multiple papers (Fisher, Sheridan, Wohletz, Heiken, Schmincke)	1970's	"Drying out" of maars; especially when water source was an aquifer.
Renato Funiciello and colleagues, Franco Barberi and colleagues	1970's	The use of lithic clasts to interpret the dynamics of a hydrovolcanic eruption when water source is an aquifer or hydrothermal system.
Volker Lorenz	1974	Vesiculated tuffs; wet surges

Heiken, Heiken and Wohletz; Walker and Croasdale	1970's; early 1980's	The use of particle size and shape to interpret hydrovolcanic processes.
Mike Sheridan	1975	Rhyolitic phreatomagmatic tuff rings; Sugarloaf Mountain, Arizona
R. V. Fisher	1977	Erosion by surges
Hans-Ulrich Schmincke	1977	Comprehensive work in the Eifel; variations in activity with variations in water access. Suggestion that "Vulcanian" be re-defined as phreatomagmatic.
Links established with experiments, computer modeling	_____	**—Move on to talks by Wohletz, Ort, and White**