Journal of Volcanology and Geothermal Research, 17 (1983) 1—29
Elsevier Science Publishers B.V., Amsterdam — Printed in The Netherlands

HYDROVOLCANISM: BASIC CONSIDERATIONS AND REVIEW

MICHAEL F. SHERIDAN[1] and KENNETH H. WOHLETZ[2]

[1] Department of Geology, Arizona State University, Tempe, AZ 85287 (U.S.A.)
[2] ESS1, Los Alamos National Laboratory, Los Alamos, NM 87545 (U.S.A.)

(Received December 7, 1982; revised and accepted April 9, 1983)

ABSTRACT

Sheridan, M.F. and Wohletz, K.H., 1983. Hydrovolcanism: basic considerations and review. In: M.F. Sheridan and F. Barberi (Editors), Explosive Volcanism. J. Volcanol. Geotherm. Res., 17: 1-29.

Hydrovolcanism refers to natural phenomena produced by the interaction of magma or magmatic heat with an external source of water, such as a surface body or an aquifer. Hydroexplosions range from relatively small single events to devastating explosive eruptive sequences. Fuel-coolant interaction (FCI) serves as a model for understanding similar natural explosive processes. This phenomena occurs with magmas of all compositions.

Experiments have determined that the optimal mass mixing ratio of water to basaltic melt for efficient conversion of thermal energy into mechanical energy is in the range of 0.1 to 0.3. For experiments near this optimum mixture, the grain-size of explosion products is always fine (less than 50 μm). The particles generated are much larger (greater than 1-10 mm) for explosions at relatively low or high ratios. Both natural and experimental pyroclasts produced by hydroexplosions have characteristic morphologies and surface textures. SEM micrographs show that blocky, equant grain shapes dominate. Glassy clasts formed from fluid magma have low vesicularity, thick bubble walls, and drop-like form. Microcystalline essential clasts result from chilling of magma during or shortly following explosive mixing. Crystals commonly exhibit perfect faces with patches of adhering glass or large cleavage surfaces. Edge modification and rounding of pyroclasts is slight to moderate. Grain surface alteration (pitting and secondary mineral overgrowths) are a function of the initial water to melt ratio as well as age. Deposits are typically fine-grained and moderately sorted, having distinctive size distributions compared with those of fall and flow origin.

Hydrovolcanic processes occur at volcanoes of all sizes ranging from small phreatic craters to huge calderas. The most common hydrovolcanic edifice is either a tuff ring or a tuff cone, depending on whether the surges were dry (superheated steam media) or wet (condensing steam media). Hydrovolcanic

0377-0273/83/$03.00

products are also a characteristic component of eruption cycles at polygenetic volcanoes. A repeated pattern of dry to wet products (Vesuvius) or wet to dry products (Vulcano) may typify eruption cycles at many other volcanoes. Reconstruction of eruption cycles in terms of water-melt mixing is extremely useful in modeling processes and evaluating risk at active volcanoes.

INTRODUCTION

Hydrovolcanism refers to volcanic phenomena produced by the interaction of magma or magmatic heat with an external source of water, such as a surface body or an aquifer (MacDonald, 1972; Sheridan and Wohletz, 1981). Hydroexplosion (Ollier, 1974; Schmincke, 1977) is an analogous term for explosive activity caused by this process. Hydromagmatic processes could even occur within deep (a few km) hydrothermal zones related to plutonic bodies. Stable isotope studies are important for determining the origin of water concentrated in explosive products erupted from the tops of large magma chambers (Forrester and Taylor, 1972; Kalamarides, 1982). This is especially true for silicic, caldera-forming eruptions (Christiansen and Blank, 1972; Lipman and Friedman, 1975; Hildreth, 1981).

Surficial hydroexplosions range from relatively small phreatic events, through common base-surge phenomena, up to devastating eruptions like the 1982 eruption of El Chichon. They may consist of a single explosion that opens a vent, hydrovolcanic pulses interspersed with purely magmatic activity, or a long series of steam-and-ash jets typical of sustained eruptions. Because water is plentiful near the surface of the earth, its relationship to erupting magma must play an important role in volcanism. For this reason it is useful to consider volcanism within a continuum that ranges from purely magmatic processes as one end member to steam eruptions at the other.

The controls of hydroexplosions are poorly understood at the present. Because the observed contact of magma with water at the earth's surface or beneath the sea does not always lead to explosive activity (e.g., Moore, 1975; Shepherd and Sigurdsson, 1982), some may be skeptical of the potential explosive energy of such a system. A general model and specific definitions for hydrovolcanic phenomena are lacking. Continuing experiments on water-melt interaction and careful observations of hydromagmatic eruptions and products should eventually lead to a better working model.

Hydrovolcanism affects all shallow (< 200 m) subaquatic volcanoes and most subaerial vents. Tuff cones and tuff rings (Heiken, 1971; Wohletz and Sheridan, 1983), which both result from this process, are second in global abundance only to scoria cones among pyroclastic vents (Green and Short, 1971). Hydrovolcanic explosions are also common activity on stratovolcanoes and calderas.

ENVIRONMENTS OF HYDROVOLCANISM

Hydrovolcanism encompasses all environments where the intermixing of water and magma produce explosive volcanic phenomena or extensive brecciation of rock and magma. Some of the important specific environments include: deep submarine, shallow submarine, littoral, lacustrine, phreatic, and subglacial. Hydromagmatism refers to a general process rather than to a specific event type or class of geologic situation. In many cases the specific hydrologic environment leading to an hydroexplosion is difficult to determine from surficial geologic data. However, the deposits may provide good evidence for the interaction of external water with melt. We concur with Schmincke (1977) that terms like hydromagmatism, hydrovolcanism, or hydroexplosion are preferred for situations where a strong interaction of water and magma can be proven, regardless of whether or not the source of the external water is known.

The subaqueous environment includes all activity beneath a standing body of water. Products from this environment have been termed subaquatic (Sigvaldason, 1968) or aquagene (Carlisle, 1963). Included in this category are submarine, littoral, sublacustrine, or other specific cases.

Submarine events (Bonatti, 1967) occur within deep (greater than 200 m) saline water (Honnorez and Kirst, 1975). The most common occurrence for this type of volcanism is at oceanic spreading centers and on large submarine volcanoes that form flat-topped guyots (Cotton, 1969) consisting of pillow basalts and hyaloclastites.

Littoral refers to near shore and shallow (less than 200 m) subaqueous activity (Wentworth, 1938). Common constructs include littoral cones that form where lava enters the sea (Moore and Ault, 1965; Fisher, 1968) or tuff cones near the shoreline such as Diamond Head and Koko craters in Hawaii (Wentworth and Winchell, 1947). Pseudocraters, such as those at Myvatn in Iceland, represent a type of littoral activity where a lava flowed into a fresh-water lake (Rittman, 1938).

Phreatic (Greek word for well, see Macdonald, 1972) refers to the eruption from the phreatic zone (ground water) of vaporized water and solid materials without juvenile clasts (Ollier, 1974). Included in this category are hydrothermal explosion craters such as those at Yellowstone National Park (Muffler et al., 1971) or New Zealand (Nairn and Wirdadiradja, 1980) which were produced by steam explosions at the top of hydrothermal systems. The deposits generally consist of massive explosion breccias that contain hydrothermally altered blocks in a clay matrix.

The term phreatomagmatic was used by Stearns and MacDonald (1946) in reference to explosions resulting from the conversion of groundwater to steam by ascending magma. It has been used for shallow lakes and submarine explosions as well. Because this is a common type of terrestrial hydrovolcanic environment, this term frequently occurs in the literature. The products are water, steam, brecciated country-rock, and must include juvenile clasts. Tuffs with a wide range of bedding structures are the common products.

Subglacial phenomena occur where magma is erupted beneath a glacier (Noe-Nygaard, 1940). In addition to deposits from massive floods (jökullaups), thick accumulations of pillow basalts, pillow breccias, massive palagonite tuffs, and stratified palagonite tuffs construct table mountains (stapi) or ridges (mobergs) above their vents (Sigvaldason, 1968; Ollier, 1974).

A MODEL FOR WATER-MELT INTERACTION

The relationship of explosive energy to crater size and particle velocity (and hence distribution of ejecta fragments) is generally expressed in terms of scaling laws. Considerable effort has been directed toward this problem with respect to large planetary impacts (Gault et al., 1963; Stöffler et al., 1975; Oberbeck, 1975). However, explosive phenomena occur over a wide range of time scales. Because volcanic explosions take place at a slower rate than thermochemical or thermonuclear explosions, scaling laws developed for hypervelocity impacts cannot be directly used to calculate the explosive energy from volcanic crater size or distribution of products. Perhaps dimensional analysis of theoretical scaling laws (Housen et al., 1983) will eventually prove appropriate for volcanic explosions. An alternative method is to extrapolate data from water/melt experiments (Wohletz and McQueen, 1981; Wohletz and Sheridan, 1982) to the scale of volcanic hydroexplosions.

NATURE OF THE PHYSICAL PHENOMENA

Hydrovolcanic eruptions can be considered to be the natural equivalent of a class of physical processes termed fuel-coolant interactions (FCI) by investigators of large industrial explosions. See Colgate and Sigurgeirsson, (1973) and Peckover et al., (1973) for applications of this theory to volcanic phenomena. FCI involves the contact of two fluids, the fuel having a temperature above the boiling point of the coolant (Board et al., 1974; Buchannan, 1974; Board and Hall, 1975; Frohlich et al., 1976; Drumheller, 1979; Corradini, 1981). The interaction generally results in vaporization of the coolant and chilling or quenching of the fuel. This process has attracted considerable interest because the vaporization often occurs at explosive rates. Examples from industry include destructive explosions at foundries where molten metal accidentally contacts water. Recent investigations have attempted to predict conditions that would lead to an FCI in the event of a nuclear core meltdown. These studies (Sandia Laboratories, 1975) were conducted to determine the controlling factors of FCI so that nuclear plants can be designed to prevent explosions.

An explosive FCI rapidly converts thermal energy to mechanical energy with a heat transfer rate greatly in excess of normal boiling by several orders of magnitude (Witte et al., 1970). The rapid vaporization of large volumes of water by magma in volcanic regimes and consequent expansion results in explosive yields that can reach one-quarter to one-third that of an equivalent mass of TNT.

The process of rapid heat transfer is periodic: pulses are separated by millisecond or shorter intervals. Initially a small volume of water is vaporized due to contact with the melt. At this stage the dominant effect of the vaporization energy is to fragment the melt which results in an increased surface area of contact between water and the melt (Corradini, 1981). The larger area of melt/water contact in turn promotes further vaporization of water that, through this feedback process, rapidly increases the total mechanical energy (PΔV) of the system.

When the total vaporization energy exceeds the limit of containment, the system explodes (Fig. 1) with the rapidly expanding vapors propelling the entrained melt fragments, sometimes with pieces of the containment chamber. At this stage the dominant effect of the vaporization energy is to accelerate the particles into the surrounding lower pressure space. If unmixed magma and water remain in the system after the initial explosion, a regular influx of melt and

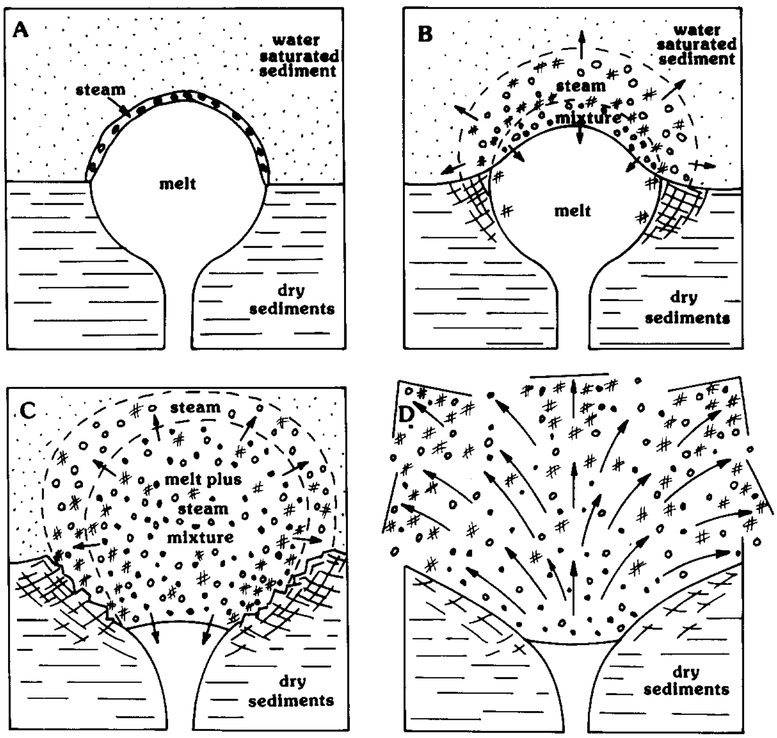


Fig. 1 Schematic diagram showing the stages of water/melt mixing within a multi-layered medium. A. Emplacement of melt into contact with water-saturated sediments. A thin vapor film develops along the contact. B. Pulsating increases in the high-pressure steam volume within the aquifer. Possible local brecciation of the country rock at this stage. C. Large-scale water/melt interaction. Mixing of country rock, steam, and melt. D. Explosive rupture of the confinement chamber.

water into the zone of mixing could lead to a sustained period of discrete explosions. Non-equilibrium thermodynamics and shock-wave physics must be considered in the analysis of this complex vaporization process. Wohletz (this volume) applies this cyclic model to develop an hypothesis on the formation of hydrovolcanic ash.

Many types of FCI phenomena have been produced under controlled conditions at Los Alamos National Laboratory (Wohletz and Sheridan, 1981; 1982; Wohletz and McQueen, 1981). The magnitude of observed explosivity varies from sporadic, pulsating ejection of large, centimeter-sized melt fragments to supersonic bursts of millimeter-sized fragments in billowing envelopes of wet condensing steam. Non-exlosive chilling and fragmentation of melt also occurred. This spectrum of experimental FCI phenomena correlates with the large-scale volcanic phenomena described as Strombolian, Surtseyan, and submarine (Walker and Croasdale, 1971).

Explosive phenomena are difficult to quantify because of the dependence of their effects on a time scale. Typically, explosive energy is scaled to a specified mass of TNT, the destructive energy of which is produced by hot, rapidly expanding gases. These expanding gases drive a shock front (the detonation wave) that causes thermal combustion of the explosive material. The rate of gas evolution depends on the velocity of the detonation wave through the material. Detonation wave velocities in high explosives are on the order of 10^3-10^4 m-s^{-1}. In contrast, the vaporization of water in FCI systems is triggered by a much slower shock wave. The propagation of such acoustic waves in systems composed of a mixture of liquid, vapor, and solids is in the range of 10^2 m-s^{-1} or less (Kieffer, 1977). Because hydrovolcanic eruptions release energy at slower rates they have less destructive potential than high explosives of equal energy. For this reason, hydrovolcanic craters can not be easily scaled to the same energy function as craters produced by high explosives.

An important aspect of FCI theory is that calculation of the explosive energy is complicated by the non-equilibrium effects of transition boiling (Buchanan and Dullforce, 1973) and superheating (Reid, 1976). These latter factors are strongly influenced by the geometry of the contact, containment pressure in the mixing zone, mass ratio of water to melt, and the temperature difference between the water and melt.

The main physical steps in an FCI cycle are summarized below:

(1) Initial contact of melt with water.

(2) Creation of a semi-insulating, superheated vapor film along the contact of the two fluids.

(3) Repeated collapse and expansion of the vapor film due to a complex balance of kinetic energy between the surrounding liquid and the film.

(4) Progressive fragmentation of the melt surface due to the kinetic energy generated by collapse and expansion of the vapor film.

(5) Increased surface contact area due to melt fragmentation and mixing with water.

(6) Increased conductive heat transfer rates with concurrent increases in film mass and energy.

(7) Vapor explosion if vaporization energy increases beyond the confinement strength (which includes surface tension effects).

At stage 7 the system can cycle back to stage 1, provided that more water and melt are available for mixing. The collapse of the steam envelopes in stage 3, which is required for explosive mixing, occurs on a time-scale of milliseconds. This collapse is inhibited by the presence of a non-condensible gas or solidification of the melt (Corradini, 1981).

DISCUSSION OF NATURAL HYDROVOLCANIC PHENOMENA

Hydroexplosions are characterized by the production of great quantities of steam and fragmented magma that are ejected from the vent in a series of eruptive pulses. Jaggar (1949) was an early advocate of the idea that many volcanic eruptions were the result of rapid vaporization of meteoric water by magma. Since the description of the 1924 eruption of Kilauea Volcano in Hawaii (Jaggar and Finch, 1924), several eyewitness accounts of other volcanic steam explosions have provided valuable information on the nature of this phenomena. These studies include the birth of maar volcano Nilahue in Chile (Muller and Vehl, 1957), the eruption of Capelinhos in the Azores (Tazieff, 1958; Servico Geologicos de Portugal, 1959), the birth of Surtsey in Iceland (Thorarinsson, 1964), and the eruption of Taal Volcano in the Philippines (Moore et al., 1966). Some important observations of the above studies include:

(1) The explosions were all periodic or pulsating.

(2) Hydroexplosions occur directly after water pours into the vent.

(3) The amount of water entering the vent and the apparent depth of explosions greatly affect the manner of pyroclast ejection.

(4) Base surges are produced.

Essentially all of the classical eruption types (Mercalli, 1907), as well as some more recently recognized types (Walker, 1973), can contain at least a small hydromagmatic component. Surtseyan activity (Thorarinsson, 1964; Walker and Croasdale, 1971) is dominantly hydromagmatic, producing mainly pyroclastic surges with minor ash or lapilli falls. Vulcanian activity (Mercali and Silvestri, 1891) has recently been shown to have a strong hydrovolcanic component (Schmincke, 1977; Nairn and Self, 1978; Frazzetta et al., this volume). Large phreatoplinian explosions produce a wide dispersal of hydrovolcanic products (Self and Sparks, 1978). The Plinian activity of Vesuvius characteristically finishes with surges and lahars (Sheridan et al., 1981; Santacroce, this volume). Strombolian activity may alternate between ash-fall and cinder production and surge clouds (Walker and Croasdale, 1971), although hydrovolcanic products are not common. Even Hawaiian volcanoes occasionally emit steam blast eruptions such as the 1790 and 1924 eruptions of Kilauea (Jaggar and Finch, 1924).

The phenomenology of volcanic steam explosions is similar in many respects to that of underwater or underground chemical or nuclear explosions (Glasstone, 1962). However, repetition of explosions from the same vent complicate the analysis of volcanic eruptions. Both types of explosions produce the following physical phenomena:

(1) Ground seismic events that can fracture the country rock near the explosion.

(2) Atmospheric acoustic events that include shock waves (Nairn, 1976; Livshits and Bolkhovitinov, 1977; and Nairn and Self, 1978). Adiabatic cooling by rarefaction behind the shock, or refraction of light at the shock front (Perret, 1912), may form visible condensation fronts that move through the atmosphere away from the vent.

(3) An ejection plume dominantly composed of steam and clasts. The emission of accidental (country rock) and juvenile fragments usually excavates a crater, or enlargens a conduit/vent system, that is surrounded by an ejecta ring.

An ejecta plume can be considered to be comprised of two vertical components (Sparks and Wilson, 1976; Wilson, 1976); a gas-thrust region and a convective thrust region. In some cases an additional horizontal component forms a base surge or a pyroclastic surge. According to Wilson (1976), the gas thrust region is characterized by rapid deceleration of erupted materials whereas the convective thrust region receives a buoyant uplift due to heating of entrained air by the hot pyroclasts.

Coarse-grained ejecta in the vertical eruption plume move in dominantly ballistic trajectories, modified to various degrees by aerodynamic drag. Fine-grained ejecta, in contrast, are entrained in an expanding buoyant gas cloud. Their rise, fall, or lateral movements depend on the density of the cloud relative to the surrounding air. Clots of ejecta form ballistic jets, characteristic of buried explosions, with cypressoid or cock's tail shape. Variations in the amount of ejecta and steam in different eruption pulses are common in hydromagmatic explosions. Calculations of ejecta dynamics, based on ballistic theory, yield velocities ranging from tens to several hundreds of meters per second (Lorenz, 1970; Fudali and Melson, 1972; Nairn, 1976; Steinberg and Babenko, 1978; Self et al., 1980).

The most devastating component of a hydrovolcanic eruption plume is a base surge (Moore et al., 1966; Waters and Fisher, 1971; Moore and Sisson, 1981). Two mechanisms of surge formation from a nuclear detonation have been proposed (Young, 1965): (1) directed blast related to overturning of the crater rim during excavation, and (2) bulk subsidence of material falling out of the vertical explosion plume. A related mehcanism is the lateral movement of the entire eruption cloud due to gravitational instability (Waitt, 1981; Malin and Sheridan, 1982). As the surge moves outward the grain concentration in the basal layer increases whereas upward streaming vapors elutriate fine ash into a buoyant overriding cloud (Wohletz and Sheridan, 1979).

A poorly understood aspect of hydroexplosions is the mode of contact between magma and water. Most obvious is the direct pouring of water into an open vent or the movement of magma into a standing body of water. In other cases maar volcanoes have erupted where only ground water was present. However, water-saturated country rock generally contains insufficient volatiles in the pore spaces for the maximum explosive mixing ratios (30 to 50 percent).

Many maar volcanoes occur along faults, suggesting that the rise of magma into a fracture-controlled zone in the aquifer may upset the hydrologic conditions sufficiently to cause a hydroexplosion. Consider a tabular-shaped body of magma moving upward through a system of fissures. Hydrostatic pressure could drive the water into the dilated fault zone during the intrusion. Water within the system may be locally heated enough to initiate small vapor explosions. Such explosions near the surface could fracture the country rock and excavate a crater increasing the magma-water contact area. Eruptions that begin this way first eject an explosion breccia composed of dominantly fragmented country rock with subordinant juvenile material (Kienle et al., 1980; Wohletz and Sheridan, 1983). As the size of the mixing zone grows, the eruptive energy ($P\Delta V$) progressively increases due to the greater volumes of water contacting the magma. By this process, eruptions may evolve from a Strombolian type with a low rate of transfer of thermal to mechanical energy to a Surtseyan type with a high efficiency of energy transfer.

A relatively deep (a few kilometers) magma chamber may also experience a sudden influx of water, as is the case for Plinian eruptions of Vesuvius (Sheridan et al., 1981; Santacroce, this volume). This could occur when pressure within the chamber becomes less than hydrostatic pressure in the aquifer toward the end of the Plinian stage, causing implosion of the chamber roof and walls. The pore water in the surrounding rocks would be at high pressures, as expected in low permeability zones adjacent to a heat source (Delaney, 1982). The sudden pressure differential could cause brecciation of the chamber walls and flooding of the chamber interior with water and accidental blocks leading to hydroexplosions.

Water could also be gradually excluded from the erupting magma interface during the course of the eruptive cycle, as was the case during the emergence of Surtsey from the sea. At the initial contact of magma with water on the sea floor, relatively quiet melt fragmentation produced pillow lavas and hyaloclastites. When the magma erupted through this pile near the surface of the sea, violent Surtseyan activity ensued. As the vent moved above sea level the activity changed to Strombolian accompanied by passive lava emission (Thorarinsson et al., 1964).

EXPERIMENTAL MODELING OF HYDROVOLCANISM

The first experiments that attempt to quantify the phenomenology of hydroexplosions were conducted at Los Alamos National Laboratory (Wohletz and McQueen, 1981; Wohletz and Sheridan, 1981; 1982). Early experiments demonstrated the feasibility of using large quantities of thermite (100 kg) to simulate magma in various configurations with water. The thermite reaction is highly exothermic:

$$3Fe_3O_4 + 8Al = 4Al_2O_3 + 9Fe + \Delta H \quad (1)$$

Thermite melt is similar to basaltic magma with respect to viscosity, density, and crystallization behavior. However, its enthalpy is about three times greater. In our experiments the excess enthalpy was used to produce a silicate melt by mixing quartz sand with the thermite (magnetite plus aluminum) in the explosion device. The controlled explosions ejected melt particles in a cloud of steam. Detailed descriptions of these experiments include characterization of the experimentally formed ash (Wohletz, this volume), experimental description and energy calculations (Wohletz and McQueen, 1981), and applications to planetary problems (Wohletz and Sheridan, 1981).

Previous and current experimental studies on FCIs have a strong bearing on the interpretation of our work. Our experimental configuration allowed quantitative measurements of the explosive phenomena using high speed cinematography of the ejecta plume as well as pressure and temperature records of the chamber. Approximately 90 kg of melt was produced in the upper chamber of the device. This melt penetrated an aluminum partition and contacted the water in the lower chamber. Vaporization caused a rapid (< 1 second) rise in the pressure sufficient to excede the burst limit (70 bars) of the vent seal. Pressure histories within the confinement vessel for various experiments showed spikes exceeding 350 bars that lasted less than one second, oscillating responses from 40 to 150 bars over periods of seconds, and a sustained response of several seconds exceeding 350 bars.

Melt fragments enclosed in a steam envelope were explosively ejected from the device. The sizes of melt fragments varied from micrometers to centimeters in diameter, ejection velocities from 10 to over 100 $m\text{-}s^{-1}$, and steam temperatures from that of condensing steam to highly superheated, expanding steam (up to at least 500°C). Particle paths followed both ballistic trajectories and turbulent, horizontally-directed flow lines.

The results of this work (Fig. 2) show that the energy of hydroexplosions depends strongly upon the mass ratio of interacting water and magma within the vent, as well as the confining pressure and geometry of the contact. Explosive efficiency is manifested by the fine-grained ejecta, superheated steam, and surging flow of materials from the orifice. Maximum explosivity for thermite experiments, measured as the conversion of efficiency of thermal to mechanical energy, occurs for ratios between 0.3 and 0.7. Because of the difference in enthalpy, these values correspond to ratios between 0.1 and 0.3 for basaltic magmas.

CHARACTERISTICS OF THE GRAINS

Clast Grain size

Abundant grain size data on surge deposits has been collected in recent

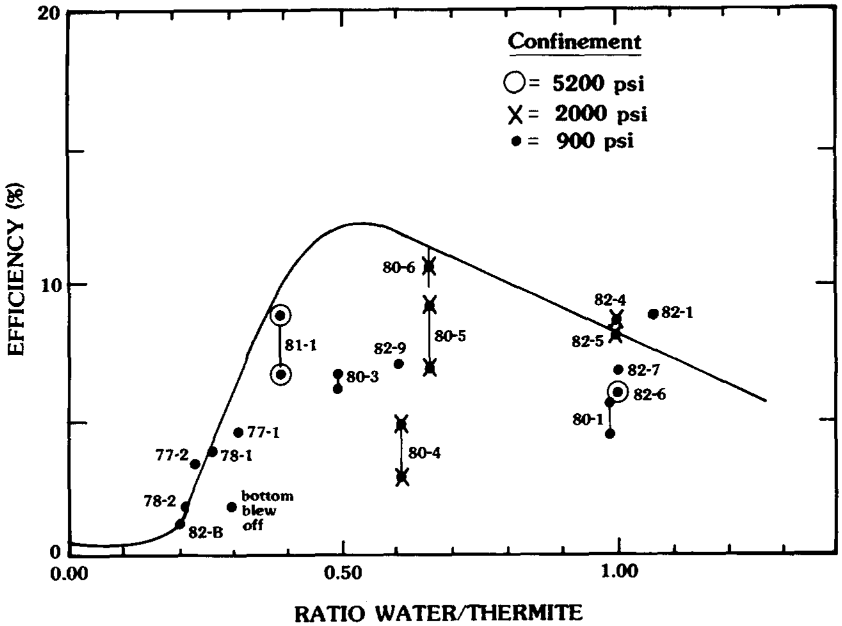


Fig. 2 Efficiency vs. water/melt ratio.

years, including information from the Azores (Walker and Croasdale, 1971), California (Crowe and Fisher, 1973), Arizona (Sheridan and Updike, 1975), southwestern U.S.A. (Wohletz and Sheridan, 1979), Idaho (Womer and Greeley, 1980), Sabatini, Italy (De Rita, et al., 1982), Mount St. Helens, Washington (Hoblitt et al., 1981), Japan (Yokoyama, 1982), Laacher See, Germany (Fisher et al., this volume), and Vulcano, Italy (Frazzetta et al., this volume). Some general observations are possible from the above data.

Walker (1971) has shown that pyroclastic-flow and pyroclastic-fall deposits occupy relatively discrete fields on a plot of median size vs sorting, but sample from surge deposits fall between the fall and flow fields. However, when considered alone, pyroclastic-surge deposits occupy a relatively well-defined field on a plot of median size and sorting (Fig. 3). Where samples are selected from different depositional units for comparison (see: Hoblitt et al., 1981; Frazzetta et al., this volume; Fisher et al., this volume), there is an apparent correspondence of grain-size distribution to bedform, as pointed out by Sheridan and Updike (1975) and Wohletz and Sheridan, (1979). Grain size gradually increases from beds with accretionary lapilli, through sandwave, massive, and planar surge beds, to lapilli-fall and breccia horizons.

The correlation of grain size with eruptive and transport mechanisms is complicated by the common incorporation of two or more discrete size populations within a single bed. Thus, moment analysis of total sample populations cannot strictly represent grain-size data of surge deposits. The separation of

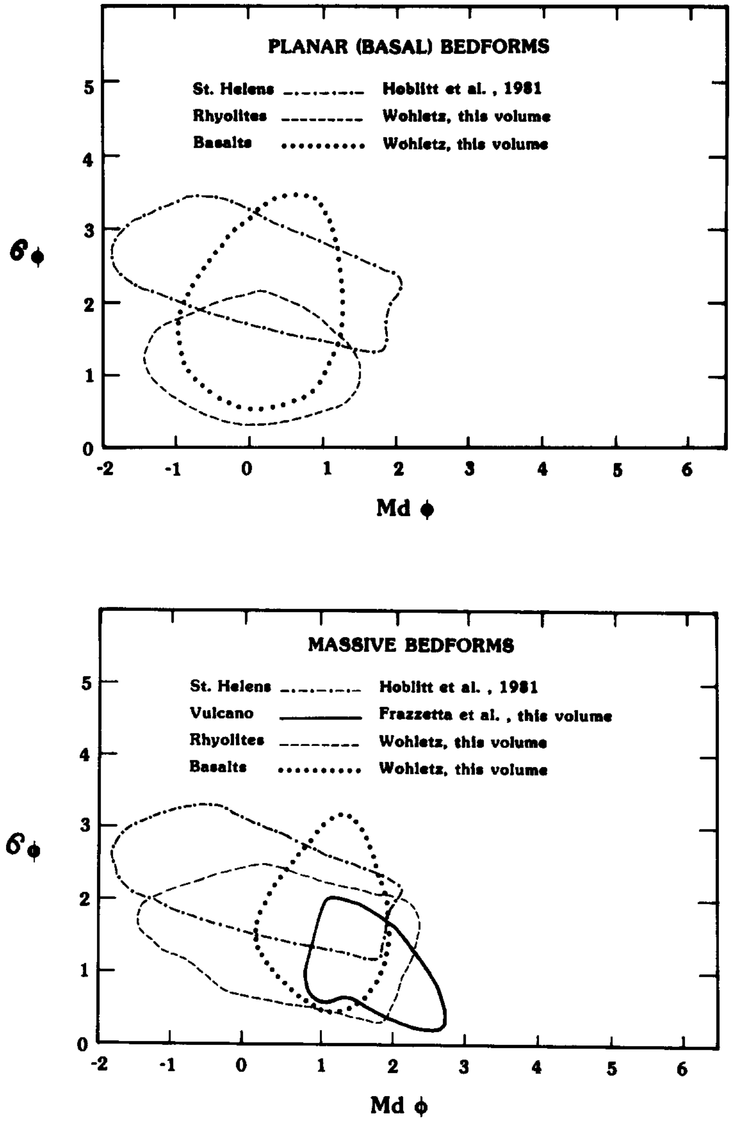
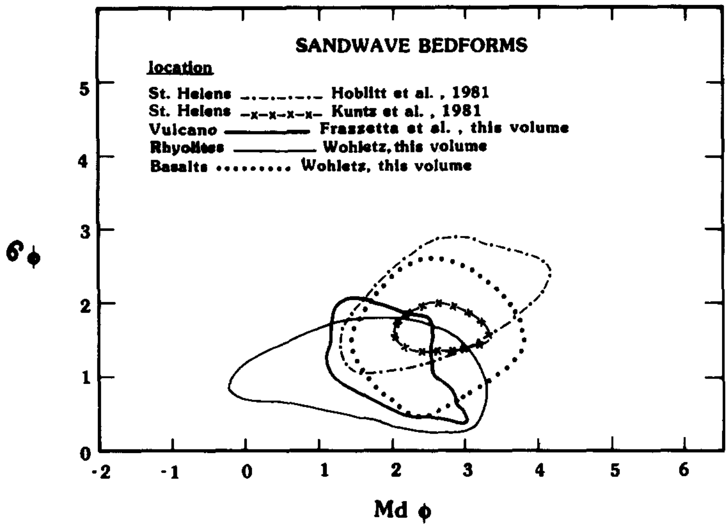
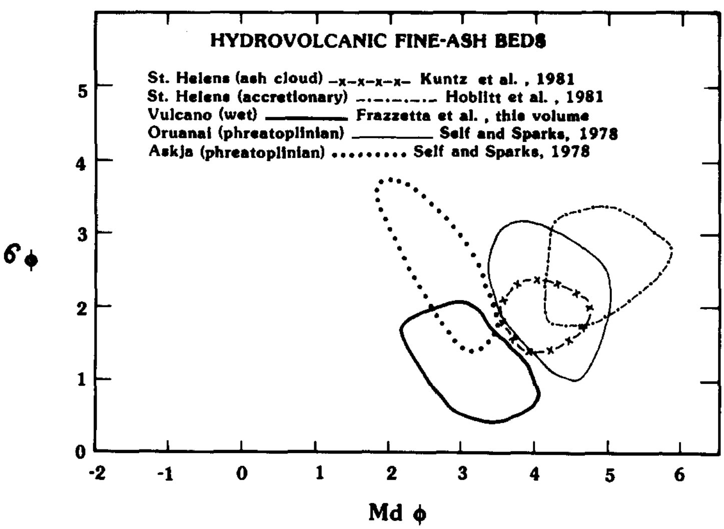


Fig. 3 Median diameter (M_ϕ) vs. sorting (σ_ϕ) for pyroclastic surge deposits.

composite particle-size data of surge deposits into their component subpopulations is a problem yet to be satisfactorily addressed. One method that has been used to distinguish subpopulations is factor analysis (Sheridan and Updike, 1975). Other techniques, such as isolating subpopulations by fitting polymodal distributions to spline functions, may also prove useful.





Particle morphology (SEM)

Hydrovolcanic deposits contain grains that have been significantly affected by a variety of processes related to their generation, transport, and alteration. Hence grain surface features may record a wealth of information about the history of the particles. Typically, clasts of different types and histories are brought together in hydrovolcanic deposits. Because of the complexity of these deposits, a methodology for the study of pyroclasts should be adopted that will optimize retrieval of this information (Sheridan and Marshall, 1982).

The distinctive shapes of pyroclasts resulting from volcanic steam explosions has been observed in many studies, most notably those of Heiken (1972; 1974). More recent studies of glassy hydrovolcanic clasts (Honnorez and Kirst, 1975; Wohletz and Krinsley, 1982) have found a wide range of particle shapes and textures than can be genetically interpreted. The common shapes of glass pyroclasts can be ascribed to varying energies and modes of contact of water with magma (Wohletz, this volume): blocky-equant (Fig. 4A), moss-like (Fig. 4B), plate-like (Fig. 4C), and drop or spherical (Fig. 4D).

Crystalline pyroclasts likewise have distinctive features (De Rita et al., 1982), although the origin of crystal surface textures is more obscure. The most common crystalline morphology are blocky grains bounded by large cleavage surfaces with smaller step fractures which are also related to cleavages (Fig. 4E). Where abundant water is present in the vent perfect crystals or perfect crystals coated with vesiculated glass are produced (Fig. 4F).

Besides the general shape of the particles, fine-scale surface features, readily observable by a scanning electron microscope, provide information on the depositional and diagenetic history of the clasts. Features related to grain collisions during transport include: adhering particles, edge modifications, grooves, scratches, step-like fractures, and dish-shaped fractures. Alteration features include vesicle fillings, skin cracks, and microcrystalline encrustations.

Thus, particle morphology reveals not only the mechanism for initial fragmentation of the melt, but also the amount and type of abrasion that occured during transport and deposition. Even post-depositional processes, such as palagonitization and lithification, are recorded. The percentage of broken, angular ash and fused or drop-like surfaces increases (but the vesicularity decreases) with increasing amounts of external water interaction during the eruption (Wohletz and Sheridan, 1982; Wohletz and Krinsley, 1983).

Alteration of grains

Glassy pyroclasts sampled from hydrovolcanic deposits display a variety of post-depositional alteration features. Outcrops of strongly altered deposits are usually orange to tan or brown in color and well-lithified, even in young deposits. In contrast, the color of outcrops composed of unaltered materials usually indicates their original composition (i.e. black to gray for basalts; white to tan for rhyolites), and the beds are generally unconsolidated.

Chemical alteration of the glass may be due to: 1) reaction with gases or fluids in the vent or hydration and diagenesis during initial cooling of the deposit, 2) post-emplacement hydrothermal activity, or 3) subsequent groundwater reactions. These three origins can be distinguished by the respective field geometries of their alteration zones: 1) bedform dependence (wet versus dry emplacement), 2) near-vent proximity (alteration independent of bedform or

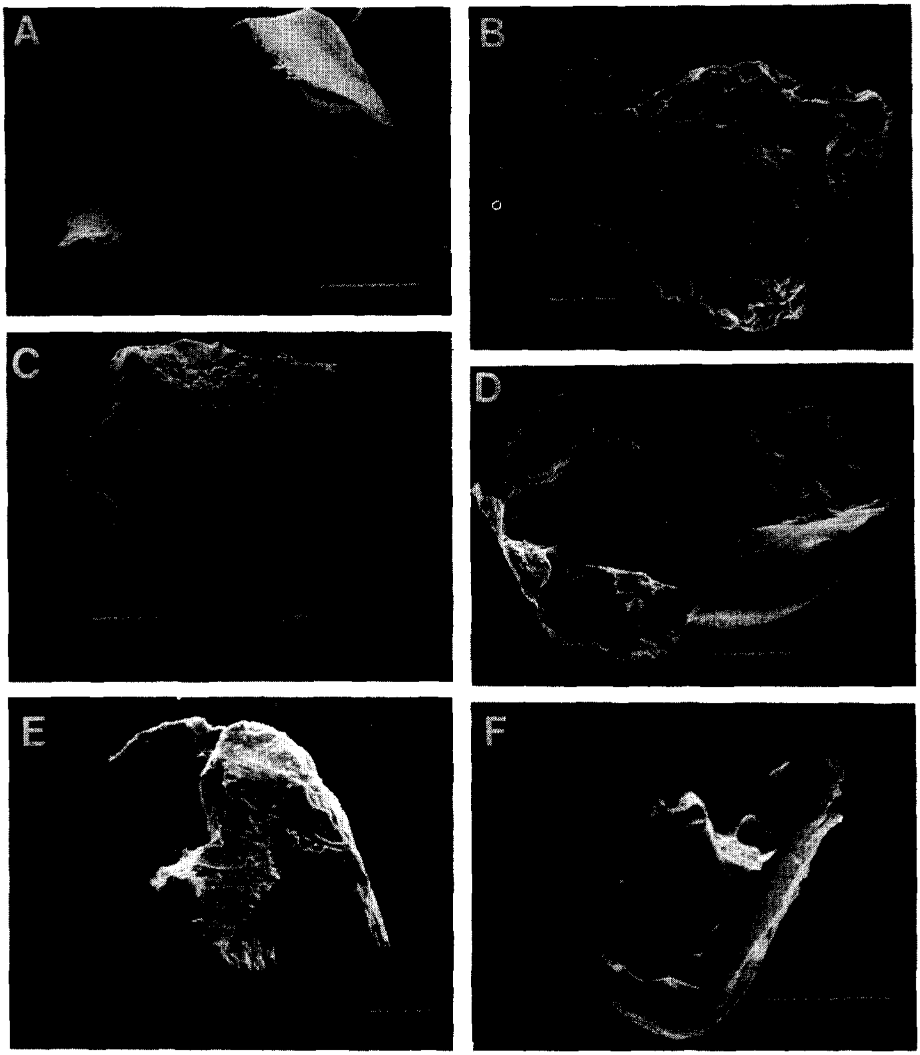


Fig. 4 SEM micrographs of typical surface features on pyroclasts from surge deposits. Glassy fragments: A. Blocky equant (Vulcano). B. Moss-like (Vulcano). C. Plate-like (Vulcano). D. Drop or fused (Taal). Crystalline pyroclasts: E. Blocky grains with large cleavage faces (Vulcano). F. Perfect crystals with adhering layer of vesiculated glass (Lipari).

horizontality), or 3) water table dependency (horizontally controlled alteration distribution). The common alteration of hydrovolcanic ash is palagonitization (Bonatti, 1965; Hay and Iijima, 1968; Honnorez, 1972; and Jakobsson, 1978) during which various elements are mobilized and redistributed on grain surfaces as a complex intergrowth of clays and zeolites.

Observation of textures on glassy particles using a scanning electron microscope reveals several stages of alteration which correspond in part to the degree of alteration and consolidation observable in outcrop. Stage 1 is marked by development of microlites of clays and zeolites within vesicle hollows or other indentations on particle surfaces. Stage 2 is noted by the formation of hydration cracks on vesicle surfaces. A thin (1 to 5 μm) hydrated skin may detach from the underlying glass between these cracks yielding an appearance similar to surface dessication textures in mud. Stage 3 alteration is characterized by the presence of hydration cracks over much of the particle surface. Overgrowths of palagonite as crusts or aggregates of fine-grained, crystalline or non-crystalline material are also abundant in this stage of alteration.

The surface chemistry of grains determined during electron microbeam analysis is also useful for distinguishing fresh and altered surfaces. Energy dispersive spectral analysis (EDS) provides adequate quantitative data for assessment of the degree of alteration. Chemical signatures change with the inferred degree of melt-water interaction from samples of tuff cone and tuff ring deposits. Strong systematic variations in iron, silica, alumina, sodium and potassium are noted between fresh and altered glass. As silica increases for these natural samples, the other elements decrease. A similar trend was found to be related to the water content of experimentally produced palagonite (Furness, 1975).

DEPOSITS

Deposits formed by hydromagmatic eruptions display a great variety of textures and structures because of the wide range of environments and explosivity of the water/melt interaction. These deposits contain primary ejecta that range from primitive to highly-evolved magma compositions. Their vents are associated with all types of feeding systems. The unifying factor for all is the contact of external water with magma that leads to fragmentation and dispersal of the clasts.

The main hydromagmatic eruption phenomena include Surtseyan, phreatoplinian, pyroclastic surge and fall, and subaqueous (including subglacial) processes. Surge phenomena (Young, 1965; Moore, 1967; Fisher and Waters, 1970) result from a range of conditions (Sheridan and Wohletz, 1981) including dry (superheated steam media) to wet (condensing steam media). The basal part of a surge may be dilute near the vent but grain concentration increases with runout distance (Wohletz and Sheridan, 1979; Fisher, 1979). The overriding cloud is generally dilute and very hot in the dry surges. Subaqueous clastic eruptions produce a mixture of particles within a water matrix that is transported at near ambient conditions. The particle concentration within such flows is extremely variable ranging from dense to dilute. Pyroclasts from phreatoplinian explosions (Self and Sparks, 1978; Self, this volume) rise in dilute, buoyant plumes to great heights in the atmosphere. The broad dispersal of clasts by lateral wind currents produces deposits at nearly ambient temperatures.

Their deposits form widespread, thin sheets of fine ash. Proximal deposits range from surge beds to stratified ash and layers with accretionary lapilli are common. Climbing megaripples and inversely graded planar stratification have been reported. The deposits show little downwind decrease in median grain size. The known vents are silicic calderas which contain lakes, the suspected source of external water.

Dry-surge deposits form thin sheets composed of unconsolidated, well-stratified deposits (Fisher and Waters, 1970). The three common types of bedforms (sandwave, massive, and planar) have a stochastic relationship that allows the definition of specific facies that are related to the distance from the source (Wohletz and Sheridan, 1979). On steep slopes proximal to the vent typical sandwave beds may be cut by U-shaped channels (Fisher, 1977) that are filled by massive density-flow deposits. Near-vent explosion breccias may cover large impact sags formed by deformation of underlying plastic layers. Many large blocks, however, are carried by the surge currents and are matrix-supported with no underlying depressions. Beds with abundant accretionary lapilli extend to medial distances where lensoid massive beds are common. Distal planar beds lack cross-stratification, but display reverse grading due to their emplacement by grain flow.

Wet surge beds typically form thick near-vent, accumulations that are strongly indurated by secondary minerals formed in the warm damp ash shortly after deposition. The beds are generally thick, massive to planar types with indistinct stratification. Mudflow and sheetwash deposits are common. Large-scale slumps and megaripples due to post-deposition deformation are common on steep slopes. Beds of vesiculated tuffs (Lorenz, 1974) and accretionary lapilli occur in most deposits. Layers plastered onto cliff faces (Heiken, 1971), trees or buildings (Waters and Fisher, 1971) attest to the cohesion provided by condensed water on grain surfaces.

Subglacial and subaqueous aquagene deposits (Carlisle, 1963) have some textural features in common with surge beds. Shallow water deposits are generally well-stratified and consist of massive and cross-stratified deposits of sand-sized clasts (Sigvaldason, 1968). The size of the fragments increases in deeper water with the progressive appearance of pillow lavas. Channels formed by turbidity flows of breccia are common on the flanks of larger volcanoes. Foreset breccia layers deposited at the angle of repose may surround some vents. Fluidization pipes and hydrobreccias are common in the vicinity of feeder dikes within volcanic constructs. Thick sections of relatively uniform hyaloclastite may form in some cases, and in others pillows may be supported in a palagonite matrix. Deposits of acid composition occur in both the submarine

(Pichler, 1965) and subglacial (Furnes et al., 1980) environments, although mafic deposits are much more common.

TYPES OF VOLCANOES AND SCALE OF THE PHENOMENA

Tuff rings and tuff cones (Heiken, 1971; Macdonald, 1972) are the most common landforms created by hydroexplosions. An understanding of these monogenetic volcanoes forms the basis for the interpretation of more complex eruption cycles in polygenetic volcanoes. Tuff rings have low topographic profiles and gentle external slopes whereas tuff cones have high profiles and steep outer slopes (Wohletz and Sheridan, 1983). Both are small volcanoes (less than 5 km diameter) and contain relatively large craters (Darwin, 1844). If the floors extend below the original ground surface they may be called maars (Ollier, 1967; Lorenz et al., 1970; 1973). Tuff rings are more commonly associated with maars than tuff cones. Their craters generally broaden and deepen as eruption progresses, leading to collapse, slump structures, and near vertical bedding inside their topographic rims (Heiken, 1971). Strongly asymmetrical deposits may be due to a change of vent location, multiple vents with different production rates, or strong prevailing winds. Crater rims may occur along crests of beds with quaquaversal dips, parallel collapse scarps, or at the intersection of adjacent craters.

The pyroclastic deposits surrounding hydrovolcanic vents range in morphology from steep-sided (30 to 35 degrees) cinder cones with small apical craters through tuff cones with moderate slopes (25 to 30 degrees) and much larger craters to tuff rings with very gentle slopes of 2 to 15 degrees. The avalanche slopes of cinder cones are due to steep angles of repose for centimeter-sized, rough-surfaced cinders. The difference in slope of tuff rings and tuff cones is due to the cohesion of the wet ash that constructs the latter structure. In addition, tuff rings rarely have rim deposit thicknesses that exceed 50 m, whereas those for tuff cones generally exceed 100 m in thickness.

The morphology of pyroclastic deposits surrounding hydrovolcanic vents is useful in determining the nature of the eruptions that produced those deposits. Tuff-ring deposits indicate high-energy surge eruptions in which mobile clouds of pyroclasts are transported relatively far from the vent. Clasts in these deposits are fine-grained and the abundant sandwave structures indicate high-energy transport. In contrast, tuff-cone deposits generally extend less than one crater diameter from the crater rim. Their tephra are relatively coarser than those of tuff-ring deposits and lapilli- or ash-fall beds are more abundant. These indicators, as well as the strong lithification due to wetness of emplacement, suggest that tuff cones result from low-energy surge and fall eruptions.

Small craters or pits are typical features of explosive activity in fumarolic geothermal areas (Muffler et al., 1971). Phreatic explosion pits can also occur where hot pyroclastic flows cover standing bodies of water (Rowley et al., 1981; Moyer, 1982). Such craters range in diameter from a few tens of meters to greater than one kilometer. They may be surrounded by a tuff cone, tuff ring, or a thin blanket of non-juvenile explosion breccia. Pseudocraters may also form by the explosive vaporization of water trapped beneath lava flows that enter water as those near Myvatn, Iceland (Rittman, 1938). These features are generally less than several hundred meters in size.

Subglacial volcanoes, common in Iceland (Walker and Blake, 1965) and Antarctica (LeMasurier, 1972), are steep-sided mountains composed of pillow lavas, pillow breccias, hyaloclastite breccias, and bedded tuffs similar in appearance to surge deposits (Jones, 1966; Sigvaldason, 1968). Some have flat-topped surfaces, with or without a vent cone, due to subaerial extrusion of lava above the level of the ice (Walker, 1965). Subaqueous volcanoes like Surtsey have an analogous structure and morphology (Kjartansson, 1966), as supported by the geologic descriptions of guyots (Christiansen and Gilbert, 1964; Moore and Fiske, 1969; Bonatti and Tazieff, 1970).

Hydrovolcanic phenomena are common at polygenetic volcanoes as well, although they seldom produce such distinctive landforms as with the monogenetic types. Their expression varies from occasional steam-blast explosions, such as the 1790 and 1924 events at Hawaii (Jaggar and Finch, 1924), to regular incorporation into the pattern of activity, as with the Plinian eruptions at Vesuvius (Sheridan et al., 1981; Santacroce, this volume) and Vulcano (Frazzetta et al., this volume). Because their deposits are relatively thin and similar in appearance to some water-laid tuffs, hydrovolcanic deposits have not been widely recognized on large volcanic structures. However, they merit much more attention because of their significance for volcanic risk evaluation and for interpretation of the role of external water in the general behavior of the volcano.

These hydrovolcanic landforms demonstrate the variety of eruption phenomena that results from the interaction of water and magma. Because water is abundant near the surface of the earth, hydrovolcanism is a likely occurrence at most volcanoes. The interaction of water and magma is also common in the subsurface, as evidenced by the brecciation of dikes and intrusions (Delaney and Pollard, 1981). Peperite (Macdonald, 1939) is produced by the shallow injection of basaltic magma into muds (Williams and McBirney, 1979).

HYDROVOLCANIC CYCLES

Hydrovolcanic phenomena occur in such a regular pattern at some volcanoes that they can be integrated into typical eruptive cycles. The interface of magmatic and hydrologic systems at central volcanoes remains fairly constant over long periods of time. A volcanic cycle can be considered as a sequence of events that follows a recognizable pattern with definable starting and ending points. Cycles may represent a period of days, as with the Plinian activity of Vesuvius, or a few centuries, as with the Fossa activity at Vulcano (Frazzetta et al., this volume). For many volcanoes a sufficient repose period exists between cycles for a soil horizon to develop at the boundary. At some central volcanoes a similar cycle may repeat several times throughout their history. Other volcanoes may exhibit an alternation of cycle types throughout its history. In general, a cycle follows a predictable pattern to its close unless the volcano-tectonic situation changes. In the case of volcanoes with a strong tendency for hydrovolcanic involvement, a cycle may record either an increase in interaction of magma with external water or a decrease in this process.

The A.D. 79 Plinian eruption of Vesuvius (Sheridan et al., 1981; Sigurdsson et al., 1982; Santacroce, this volume) is an excellent example of an eruption cycle that exhibits an increase in the hydromagmatic component with time. The eruption starts with a Plinian-fall layer overlain by stratified-fall deposits that are interspersed with surge beds. The next sequence of beds consists of pumice and ash flows, separated by surge horizons. The uppermost unit is a "wet" surge deposit with abundant accretionary lapilli.

The eruptions from Fossa of Vulcano (Frazzetta et al., this volume) represent typical examples of the reverse type of cycle that shows a decrease in hydromagmatic component with time. These cycles begin with lahars or wet surges and proceed to dry surge beds. These deposits are overlain by pumice-fall beds and the sequence is capped by a cycle-ending lava flow.

The above examples suggest that several textural indicators can be placed in an order of increasing interaction of external water with magma in order to define the progress of the cycle: lava flows, lapilli fall, stratified lapilli fall, dry surge with cross-laminations, accretionary lapilli, vesiculated tuffs, wet surge, lahar, pillows, and hyaloclastite. This data can be combined with the experimental results of Wohletz and McQueen (1981) to map the water to melt ratios during the cycle. An example of this technique for the Vesuvius and Vulcano types is shown in Fig. 5.

SUMMARY

Hydrovolcanism is a common natural phenomena that occurs in every volcanic setting. Its role is generally underestimated in volcanic eruptions. Fuel coolant interactions (FCI) are an industrial analog of natural hydroexplosions that serve as a model for hydrovolcanism. The explosive mixing of water and magma produces very fine melt fragmentation (10 to 50 μm). Experiments at Los Alamos National Laboratory have duplicated most of the phenomenology associated with hydrovolcanic explosions. The optimum mixing ratio of water to basaltic magma for efficient transfer of thermal energy to mechanical work is 0.1 to 0.3 mass fraction. Pyroclast morphology, bedding structures, and deposit morphology are distinctive and can be used to estimate the water to melt mixing ratio of the eruption (Fig. 6). The change in the character of the deposit throughout an eruption cycle can be used to interpret the behavior of the volcano and to evaluate potential hazards related to future activity.

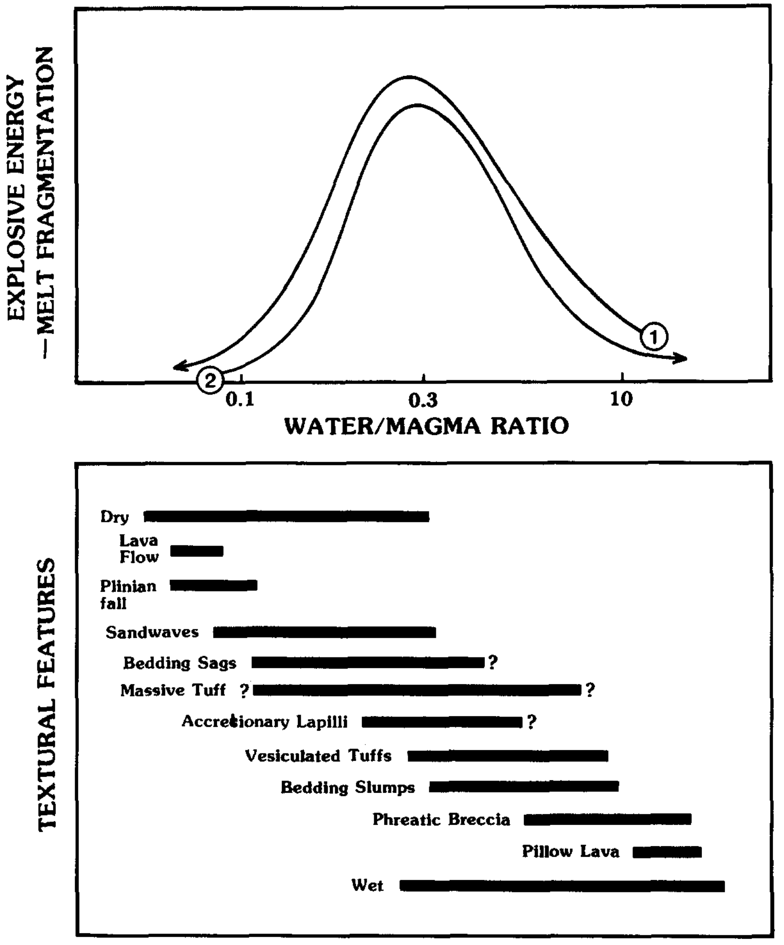


Fig. 5 Plinian (2) vs. Vulcanian (1) eruption cycles plotted on an energetics diagram. Various textures associated with water/melt ratio ranges shown below.

This paper has benefited through extended field discussions of hydrovolcanic products at Italian volcanoes with a number of people, especially: Franco Barberi, Mauro Rosi, Roberto Santacroce, Luigi La Volpe, Giovanni Frazzetta, Donatella De Rita, Gian Zuffa, Rosana De Rosa, Renato Funiciello, and Tom Moyer. The experimental work at Los Alamos National Laboratory was done with the cooperation of Robert McQueen. The SEM work was done at Los Alamos National Laboratory and Arizona State University. We would like to thank Stephen Self and Grant Heiken for critically reading the manuscript. The research was partially supported by NASA grant NAGW-245 and NSF grant INT-8200856.

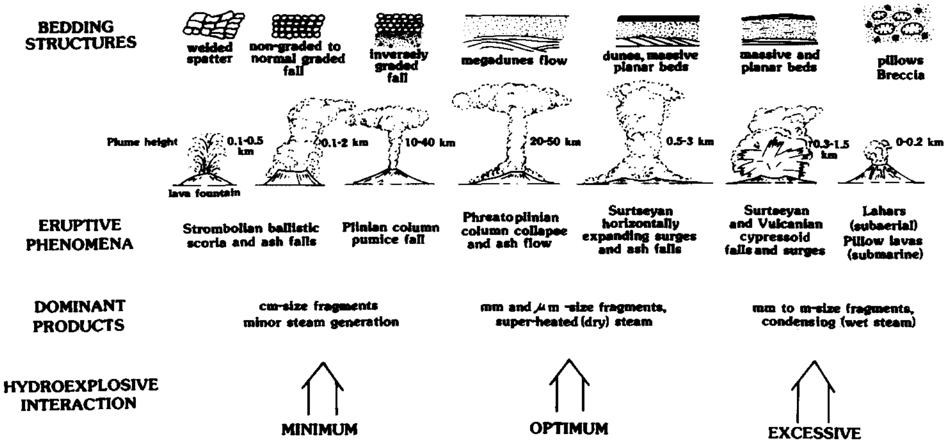


Fig. 6 The systematics of hydrovolcanic activity.

REFERENCES

Board, S.J. and Hall, R.W., 1975. Thermal explosions at molten tin/water interfaces. In: J.R. Okendon and W.R. Hodgkins (Editors), Moving boundary problems in heat flow diffusion. Clarendon Press, Oxford, pp. 259-269.

Board, S.J., Hall, R.W., and Hall, R.S., 1974. Fragmentation in thermal explosions. Inst. J. Heat Mass Transfer, 17: 331-339.

Bonatti, E., 1965. Palagonite, hyaloclastites and alteration of volcanic glass in the ocean. Bull. Volcanol., 28: 1-15.

Bonatti, E., 1967. Mechanisms of deep sea volcanism in the South Pacific. In: P.H. Ableson (Editor), Researches in Geochemistry, 2: 453-491. John Wiley & Sons, New York, N.Y.

Bonatti, E. and Tazieff, H., 1970. Exposed guyot from the Afar Rift, Ethiopia. Science, 168: 1087-1089.

Buchanan, D.S., 1974. A model for fuel-coolant interactions. J. Phys. D.: Appl. Phys., 7: 1441-1457.

Buchanan, D.S. and Dullforce, T.A., 1973. Mechanism for vapor explosions. Nature, 245: 32-34.

Carlisle, D., 1963. Pillow breccias and their aquagene tuffs: Quandra Island, British Columbia. Amer. J. Sci., 71: 48-71.

Christiansen, M.N. and Gilbert, C.M., 1964. Basaltic cone suggests constructional origin of some guyots. Science, 143: 240-242.

Christiansen, R.L. and Blank, H.R., Jr., 1972, Volcanic stratigraphy of the Quaternary rhyolite plateau in Yellowstone National Park. U.S. Geol. Survey Prof. Paper, 729-B, 18 pp.

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

Corradini, M.L., 1981. Analysis and modelling of steam explosion experiments. Sandia Natl. Labs., SAND80-2131, NUREG/CR-2072, pp. 114.

Cotton, C.A., 1969. The pedestals of oceanic volcanic islands. Geol. Soc. Amer. Bull., 80: 749-760.

Crowe, B.M. 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.R., 1844. Geologic observations on the volcanic islands visited during the voyage of HMS Beagle, with brief notes on the geology of Australia and the Cape of Good Hope, being the second part of the geology of the voyage of the Beagle. London.

Delaney, P.T., 1982. Rapid intrusion of magma into wet rock: groundwater flow due to pore-pressure increases. Jour. Geophys. Res., 87: 7739-7756.

Delaney, P.T. and Pollard, D.D., 1981. Deformation of host rock and flow of magma during growth of minette dikes and breccia-bearing intrusions near Ship Rock, New Mexico. U.S. Geol. Survey Prof. Paper 1202, 61 pp.

De Rita, D., Sheridan, M.F., and Marshall, J.R., 1983. SEM surface textural analysis of phenocrysts from pyroclastic deposits at Baccano and Sacrofano volcanoes, Latium, Italy. In: W.B. Whalley and D.H. Krinsley (Editors), SEM in Geology. Geoabstracts, Norwich, England, in press.

Drumheller, D.S., 1979. The initiation of melt fragmentation in fuel-coolant interactions. Nucl. Sci. Eng., 72: 347-356.

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

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

Fisher, R.V., 1979. Models for pyroclastic surges and pyroclastic flows. J. Volcanol. Geotherm. Res., 6: 305-318.

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

Fisher, R.V., Schmincke, H.U., and Van Bogaard, P., 1983. Origin and emplacement of a pyroclastic flow and surge unit at Laacher See, Germany. In: M.F. Sheridan and F. Barberi (Editors), Explosive Volcanism. J. Volcanol. Geotherm. Res., 17: (this volume).

Forrester, R.W. and Taylor, H.P., Jr., 1972. Oxygen and hydrogen isotope data on the interaction of meteoric ground water with a gabbro-diorite stock, San Juan Mountains, Colorado. Int. Geol. Cong., 24th, Montreal, Geochem. Proc., 10(24): 254-263.

Frazzetta, G., La Volpe, L., and Sheridan, M.F., 1983. Evolution of the Fossa cone, Vulcano. In: M.F. Sheridan and F. Barberi (Editors), Explosive Volcanism. J. Volcanol. Geotherm. Res., 17: (this volume).

Fröhlich, G., Müller, G., and Unger, G., 1976. Experiments with water and hot melts of lead. J. Non-Equilib. Thermodyn., 1: 91-103.

Fudali, A.F., and Melson, W.G., 1972. Ejecta velocities, magma chamber pressures, and kinetic energy associated with the 1968 eruption of Arenal Volcano. Bull. Volcanol., 35: 381-401.
Furnes, H., 1974. Volume relations between palagonite and authigenic minerals in hyaloclastites, and its bearing on the rate of palagonitization. Bull. Volcanol., 38: 173-186.
Furnes, H., Fridleifsson, I.B., Atkins, F.B., 1980. Subglacial volcanoes: on the formation of acid hyaloclastites. J. Volcanol. Geotherm. Res., 8: 137-160.
Gault, D.E., Shoemaker, E.M., and Moore, H.J., 1963. Spray ejecta from the lunar surface by meteoroid impact. NASA Tech. Note, D-1767.
Glasstone, S., 1950. (Editor), The effects of atomic weapons, Los Alamos, New Mexico. U.S. Atomic Energy Comm., Los Alamos Sci. Lab., N.M., 156 pp.
Green, J. and Short, N.M., 1971. Volcanic landforms and surface features, a photographic atlas and glossary. Springer-Verlag, New York, pp. 522.
Hay, R.L. and Iijima, A., 1968. Nature and origin of palagonite tuffs of the Honolulu group on Oahu, Hawaii. Geol. Soc. Amer. Mem., 116: 331-376.
Heiken, G., 1971. Tuff rings: examples from the Fort Rock-Christmas Lake Valley basin, south-central Oregon. J. Geophys. Res., 76: 5615-5626.
Heiken, G., 1972. Morphology and petrography of volcanic ashes. Geol. Soc. America Bull., 83: 1961-1988.
Heiken, G., 1974. An atlas of volcanic ash. Smithsonian Contrib. Earth Sci., no. 12, pp. 101.
Hildreth, W., 1981. Gradients in silicic magma chambers: implications for lithospheric magmatism. J. Geophys. Res., 86: 10153-10192.
Hoblitt, R.P., Miller, C.D., and Vallance, J.W., 1981. Origin and stratigraphy of the deposit produced by the May 18 directed blast. In: P.W. Lipman and D.R. Mullineaux (Editors), The 1980 eruptions of Mount St. Helens, Washington. U.S. Geol. Survey Prof. Paper 1250: 401-420.
Honnorez, J., 1972. La Palagonitisation. Vulkaninstitut Immanuel Friedllaender, No. 9. Birkhauser Verlag, Basel., 131 pp.
Honnorez, J. and Kirst, P., 1975. Submarine basaltic volcanism: morphometric parameters for discriminating hyaloclastites from hyalotuffs. Bull. Volcanol., 39: 441-465.
Housen, K.R., Schmidt, R.M., and Holsapple, K.A., 1983. Crater ejecta scaling laws: fundamental forms based on dimensional analysis. J. Geophys. Res., 88: 2485-2499.
Jaggar, T.A., 1949. Steam blast volcanic eruptions. Hawaiian Volcano Obs., 4th Spec. Rept., pp. 137.
Jaggar, T.A. and Finch, R.H., 1924. The explosive eruption of Kilauea in Hawaii, 1924. Am. J. Sci., 8: 353-374.
Jakobsson, S.P., 1978. Environmental factors controlling the palagonitization of Surtsey tephra, Iceland. Geol. Soc. Denmark Bull. 27: 91-105.

Jones, J.G., 1966. Intraglacial volcanoes of southwest Iceland and their significance in the interpretation of the form of the marine basaltic volcanoes. Nature, 212: 586-588.

Kalamarides, R.I., 1982. $\delta^{18}O$ fractionation among phases and evaluation of contamination of the Kiglapait layered intrusion. Am. Geophys. Union Trans., EOS, 63(45): 1151.

Kieffer, S.W., 1977. Sound speed in liquid-gas mixtures: water-air and water-steam. J. Geophys. Res., 82: 2895-2904.

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.

Kjartansson, G., 1966. A comparison of table mountains in Iceland and the volcanic island of Surtsey of the south coast of Iceland. Natturufraedingurinn, 36: 1-34.

Lemasurier, W.E., 1972. Volcanic record of Cenozoic glacial history of Marie Byrd Land. In: R.J. Adie (Editor), Antarctic Geology and Geophysics. Universitetsforglaget, Oslo. pp. 251-260.

Lipman, P.W. and Friedman, I., 1975. Interaction of meteoric water with magma: an oxygen-isotope study of ash-flow sheets from southern Nevada. Geol. Soc. Amer. Bull., 86: 695-702.

Livshits and Bolkhovilinov, 1977. Weak shock wave in the eruption column. Nature, 267: 420-421.

Lorenz, V.W., 1970. Origin of Hole-in-the-ground, a maar in central Oregon. Unpub. NASA Rept., 113 pp.

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

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

Lorenz, V.W., McBirney, A.R., and Williams, H., 1970. An investigation of volcanic depressions. Part III, maars, tuff rings, tuff cones, and diatremes. Houston, NASA Prog. Rept. NGR 38003012, pp. 198.

Macdonald, G.A., 1939. An intrusive peperite at San Pedro Hill, Calif. Univ. Calif. Pub., Bull. Geol. Sci., 24: 329-338.

Macdonald, G.A., 1972. Volcanoes. Prentice-Hall, Inc., N.J., 510 pp.

Malin, M.C. and Sheridan, M.F., 1982. Computer-assisted mapping of pyroclastic surges. Science, 217: 637-639.

Mercali,, G., 1907. I vulcani attivi della terra. Ulrico Hoepli, Milan, 421 pp.

Mercali, G. and Silvestri, O., 1891. Le eruzioni dell'isola di Vulcano, incominciate il 3 Agosto 1888 e terminate il 22 Marzo 1890. Relazione Scientifica, 1891 Ann. Uff. Cent. Metr. Geodin., 10(4): 1-213.

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

Moore, J.G., 1975. Mechanism of formation of pillow lava. Amer. Sci., 63: 269-277.

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

Moore, J.G. and Fiske, R.S., 1969. Volcanic substructure inferred from dredge samples and ocean bottom photographs, Hawaii. Geol. Soc. Amer. Bull., 80: 1191-1202.

Moore, J.G. and Sisson, T.W., 1981. Deposits and effects of the May 18 pyroclastic surge, In: P. W. Lipman and R. R. Mullineaux (Editors), The 1980 eruptions of Mount St. Helens. U.S. Geol. Survey Prof. Paper, 1250: 421-438.

Moore, J.G., Nakamura, K., and Alcaraz, A., 1966. The 1965 eruption of Taal Volcano. Science, 151: 995-960.

Moyer, T.C., 1982. Deposits of the phreatic-explosion pits of the 18 May, 1980 Mount St. Helens pyroclastic flows. Am. Geophys. Union. EOS, 63(45): 1141.

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

Muller, G. and Vehl, G., 1957. The birth of Nilahue, a new maar type volcano at Rinihue, Chile. Int. Geol. Cong. 25th, Mexico 1956, Rept. sec. 1, 375-396 pp.

Nairn, I.A., 1976. Atmospheric shock waves and condensation clouds from Ngauruhoe explosive eruptions. Nature, 259: 190-192.

Nairn, I.A. and Self, S., 1978. Explosive eruptions and pyroclastic avalanches from Ngauruhoe in February, 1975. J. Volcanol. Geotherm. Res., 3: 39-60.

Nairn, I.A. and Wirdadiradja, S., 1980. Late Quaternary hydrothermal explosion breccias at Kawerau geothermal field, New Zealand. Bull. Volcanol., 43: 1-14.

Noe-Nygaard, A., 1940. Subglacial volcanic activity in ancient and recent times. Folia Geogr. Danica, 1: no. 2.

Oberbeck, V.R., 1975. The role of ballistic erosion and sedimentation in the lunar stratigraphy. Rev. Geophys. Space Phys., 13: 337-362.

Ollier, C.D., 1967. Maars: their characteristics, varieties and definitions. Bull. Volcanol., 31: 232-247.

Ollier, C.D., 1974. Phreatic eruptions and maars. In: L. Civetta et al., (Editors), Physical Volcanology. Elsevier, New York, N.Y., pp. 289-310.

Peckover, R.S., Buchanan, D.J., and Ashby, D.E.T.F., 1973. Fuel-coolant interactions in submarine volcansim. Nature, 245: 307-308.

Perret, F.A., 1912. Eruptions of Vesuvius, 1906. Am. J. Sci., 4: 329-333.

Pichler, H., 1965. Acid hyaloclastites. Bull. Volcanol., 28: 293-310.

Reid, R.G., 1976. Superheated liquids. Amer. Sci., 64: 146-156.

Rittman, A., 1938. Die Vulkane am Myvatn in Nordost-Island. Bull. Volcanol., 4: 3-38.

Rowley, P.D., Kuntz, M.A., and Macleod, N.S., 1981. Pyroclastic flow deposits. In: P.W. Lipman and D.R. Mullineaux (Editors), The 1980 Eruptions of Mount St. Helens, Washington, U.S. Geol. Survey Prof. Paper, 1250: 489-512.

Sandia Laboratories, 1975. Core meltdown experimental review. SAND74-0382, pp. 472.

Santacroce, R., 1983. A general model for the behavior of the Somma-Vesuvius volcanic complex. In: M.F. Sheridan and F. Barberi (Editors), Explosive Volcanism. J. Volcanol. Geotherm. Res., 17: (this volume).

Schmincke, H.U., 1977. Phreatomagmatische Phasen in quartaren Vulkanen der Osteif. Geol. Jahrbuch., A39: 3-45.
Self, S. and Sparks, R.S.J., 1978. Characteristics of widespread pyroclastic deposits formed by the interaction of silicic magma and water. Bull. Volcanol., 41(3): 196-212.
Self, S., Kienle, J., and Huot, J.P., 1980. Ukinrek maars, Alaska, II. Deposits and formation of the 1977 craters. J. Volcanol. Geotherm. Res., 7: 39-65.
Servico Geologicos de Portugal, 1959. Le volcanisme de l'isle de Faial et l'eruption de volcan de Capelinhos. Mem. 4 (nova ser.), 100 pp.
Shepherd, J.B. and Sigurdsson, H., 1982. Mechanism of the 1979 explosive eruption of Soufriere Volcano, St. Vincent J. Volcanol. Geotherm. Res., 13: 119-130.
Sheridan, M.F. and Updike, R.G., 1975. Sugarloaf Mountain Tephra: A Pleistocene rhyolitic deposit of base-surge origin in northern Arizona. Geol. Soc. Amer. Bull., 86: 571-581.
Sheridan, M.F. and Wohletz, K.H., 1981. Hydrovolcanic explosions: the systematics of water-pyroclast equilibration. Science, 212: 1387-1389.
Sheridan, M.F. and Marshall, J.R., 1983. SEM examination of pyroclastic materials: basic considerations. Scanning Electron Microscopy, in press.
Sheridan, M.F., Barberi, F., Rosi, M., and Santacroce, R., 1981. A model for the eruptions of Vesuvius. Nature, 289: 282-285.
Sigurdsson, H., Cashdollar, S., and Sparks, R.S.J., 1982. The eruption of Vesuvius in A.D. 79: reconstruction from historical and volcanological evidence. Am. Jour. Archeol., 86: 39-51.
Sigvaldason, G., 1968. Structure and products of subaquatic volcanoes in Iceland. Contr. Mineral. and Petrol., 18: 1-16.
Sparks, R.S.J. and Wilson, L., 1976. A model for the formation of ignimbrite by gravitational column collapse. Geol. Soc. London J., 132: 441-451.
Stearns, H.T. and MacDonald, G.A., 1946. Geology and groundwater resources of the island of Hawaii. Hawaii Div. Hydrol. Bull., 67: 13-49.
Steinberg, G.S. and Babenko, J.I., 1978. Experimental velocity and density determination of volcanic gasses during eruption. J. Volcanol. Geotherm. Res., 3: 89-98.
Stöffler, D., Gault, D.E., Wedekind, J., and Polkowski, G., 1975. Experimental hypervelocity impact into quartz sand: distribution and shock metamorphism of ejecta. J. Geophys. Res., 80: 4062-4077.
Tazieff, H.K., 1958. L'eruption 1957-1958 et la tectonique de Fail (Azores). Soc. Belg. Geol. Bull., 67: 13-47.
Thorarinsson, S., 1964. Surtsey, the new island in the North Atlantic. Reykjavik, Almenna Bokofelagid, pp. 47.
Thorarinsson, S., Einarsson, T., and Sigvaldason, G., 1964. The submarine eruption of the Vestmann Islands, 1963-63. Bull. Volcanol., 27: 437-446.

Waitt, R.B., 1981. Devastating pyroclastic density flow and attendant air fall of May 18 -- stratigraphy and sedimentology of deposits. In: P.W. Lipman and D.R. Mullineaux (Editors), The 1980 Eruptions of Mount St. Helens, Washington. U.S. Geol. Survey Prof. Paper, 1250: 439-460.

Walker, G.P.L., 1971. Grain size characteristics of pyroclastic deposits. J. Geol., 79: 696-714.

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. The formation of a palagonite breccia mass beneath a valley glacier in Iceland. Quart. J. Geol. Soc. London, 122: 45-61.

Walker, G.P.L. and Croasdale, R., 1971. Characteristics of some basaltic pyroclasts. Bull. Volcanol., 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., 1938. Ash formations of the island of Hawaii. Hawaii Volcano Observ. 3rd Spec. Rept., Honolulu, Hawaiian Volcano Res. Assoc., pp. 173.

Wentworth, C.K. and Winchell, H., 1947. Koolau basalt series, Oahu, Hawaii. Geol. Soc. Amer. Bull., 58: 49-78.

Williams, H. and McBirney, A., 1979. Volcanology. Freeman, Cooper & Co., San Francisco. 397 pp.

Wilson, L., 1976. Explosive volcanic eruptions - III. Plinian eruption columns. Geophys. J. Roy. Soc., 45: 543-556.

Witte, L.C., Cox, J.E., and Bouvier, J.E., 1970. The vapor explosion. J. Metals, 22: 39-44.

Wohletz, K.H., 1983. Mechanism of hydrovolcanic pyroclast formation: Size, scanning electron microscopy, and experimental studies. In: M.F. Sheridan and F. Barberi (Editors), Explosive Volcanism. J. Volcanol. Geotherm. Res., 17: (this volume).

Wohletz, K.H. and Sheridan, M.F., 1979. A model of pyroclastic surge. In: C.E. Chapin and W.E. Elston (Editors), Ash-flow Tuffs. Geol. Soc. America Spec. Paper, 180: 177-194.

Wohletz, K.H. and Sheridan, M.F., 1981. Rampart crater ejecta: experiments and analysis of melt/water interactions. NASA Tech. Memor. 82385: 134-136.

Wohletz, K.H. and McQueen, R., 1981. Experimental hydromagmatic volcanism. Amer. Geophys. Union Trans., EOS, 62(45): 1085.

Wohletz, K.H. and Sheridan, M.F., 1982. Melt/water interactions: series II experimental design. NASA Tech. Memor., 84211: 169-171.

Wohletz, K.H. and Krinsley, D.H., 1983. Scanning electron microscopic analysis of basaltic hydromagmatic ash. In: W.B. Whaley and D.H. Krinsley (Editors), SEM in Geology. Geoabstracts Norwich, England, (in press).

Wohletz, K.H. and Sheridan, M.F., 1983. Hydrovolcanic explosions II. Tuff rings and tuff cones. Am. J. Sci., in press.

Womer, M.B., Greeley, R., and King, K.S., 1980. The geology of Split Butte -- a maar of the south-central Snake River Plain, Idaho. Bull. Volcanol., 43: 453-471.

Yokoyama, S., 1980. Base surge deposits in Japan. In: S. Self and R.S.J. Sparks (Editors), Tephra Studies. D. Reidel Pub. Co., Dordrecht, Holland. pp. 427-432.

Young, G.A., 1965. The physics of the base surge. White Oak, Maryland, U.S. Naval Ordinance Lab., NOLTR64-130, pp. 284.