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Notes

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SUMMARY: The type, relative abundance and stratigraphical relationships of volcanic rocks that comprise island volcanoes are a function of (i) depth of extrusion beneath water, (ii) magma composition, and (iii) lava–water interactions. The water depth at which explosions can occur is called the pressure compensation level (PCL) and is variable. Explosive eruptions that occur above the PCL and below sealevel can give rise to abundant hydroclastic and pyroclastic debris. Below the PCL, clastic material cannot form explosively; it forms from lava by thermal shock. The volcaniclastic products are widely dispersed in basins adjacent to extrusion sources by three principal kinds of marine transport processes. These are slides, sediment gravity flows and suspension fallout. Volcaniclastic debris can be derived in subaqueous and subaerial-to-subaqueous environments (i) directly from eruptions, (ii) from remobilization of juvenile volcaniclastics, or (iii) from epiclastic material which initially develops above sealevel.

Sediment gravity flows (fluids driven by sediment motion) exhibit the phenomenon of *flow transformation*. This term is used here for the process by which (i) sediment gravity flow behaviour changes from turbulent to laminar, or vice versa, within the body of a flow, (ii) flows separate into laminar and turbulent parts by gravity, and (iii) flows separate by turbulent mixing with ambient fluid into turbulent and laminar parts. Dominant kinds of subaqueous volcaniclastic sediment gravity flows are debris flows, hot or cold pyroclastic flows and turbidites. Fine grained material can be thrown into suspension locally during flow transformations or underwater eruptions, but thin, regionally distributed subaqueous fallout tephra is mostly derived from siliceous Plinian eruptions.

Volcaniclastic rocks in marine sequences occur in many kinds of sedimentary environments and tectonic settings (Table 1). Some are the result of eruptions on land which deliver fallout ash, lava and pyroclastic flows into water. Others are derived from underwater eruptions which extrude lava flows and volcanic fragments of various kinds, or are pyroclastic, hydroclastic or epiclastic materials reworked from land or remobilized under water. This review is mostly concerned with subaqueous sediment gravity flow deposits, because of their abundance in marine environments adjacent to volcanic regions.

The site of most voluminous volcanism is at divergent plate boundaries where basaltic sheet flows, pillow basalts and smaller amounts of pillow breccias and hyaloclastites are formed. An estimated 4–6 km^3 of such material is added each year to the Earth's crust at mid-ocean ridges (Nakamura 1974). The second important site of volcanism is at convergent plate boundaries. Basaltic and andesitic island arcs develop at converging oceanic plates; converging oceanic and continental plates give rise to dominantly andesitic to rhyolitic volcanic chains on the edges of continents. Perhaps the most varied and complex environments are the different kinds of back-arc settings. A third important site of oceanic volcanism is represented by intra-plate seamounts and ocean islands. Submarine volcaniclastic deposits are most abundant near island arcs and ocean islands because large volumes of volcaniclastic material are transported from land into the sea during eruptions and during subsequent erosion. Volcanogenic sedimentation and tectonics, an area of considerable research, is reviewed by Mitchell & Reading (1978).

Underwater eruptions

Whether underwater eruptions are effusive or explosive is determined by (i) the depth (pressure) of the water column (Fig. 1), (ii) the composition of the magma, especially amounts of volatiles, and (iii) the extent of interaction between magma and water. Subsequent transport depends upon slope, which in turn partly depends upon the growth rate of the volcano and the initial lava-to-clastic ratio of extruded products. The manner of transport also depends upon whether or not eruption columns can develop. When vapour pressure in the magma exceeds water pressure, vesiculation commences. As depth decreases, vesicles become more abundant (Moore 1965, 1970; Moore & Schilling 1973) and, at a shallow level, explosions caused by exsolution of magmatic volatiles can occur. This depth, which is variable, is here called the pressure compensation level (PCL) and 'volatile fragmentation depth' by

TABLE 1. *Simplified partitioning of environment, kind of extrusion, transport and emplacement processes and some typical pyroclastic deposits*

Environment	Eruptions	Directly from eruptions: Transport	Directly from eruptions: Emplacement	Deposits
Subaqueous (marine, lacustrine, sub-ice)	Effusive (shallow or deep)	Lava flows	Congealing flows	Massive flows; pillow lavas; hyaloclastites.
	Explosive (shallow)	Dilute suspension in water	Suspension fallout	Thin, well sorted, normally graded beds. May be reworked by bottom currents. Local occurrence near source.
		Pyroclastic flows	Suspension fallout (dilute suspensions from tops of flows).	Thin, fine grained well sorted, normally graded beds. May be reworked by bottom currents. Rest on turbidites.
			Turbulent flows from tops of mass flows.	Thin sequence of fairly well sorted beds, may be cross-bedded. Tops may be reworked by bottom currents. Rest on pyroclastic flows.
			Laminar mass flows.	Thick, poorly sorted, poorly bedded, non-welded. Mixing with water can result in lahars. Tops may be reworked by bottom currents. May contain rip-ups. Bases erosive to non-erosive.
		(Flows also develop from slump and flow of water-logged pyroclastic debris on subaqueous slopes of volcanoes to give turbidites and lahars.)		

Subaerial eruptions with subaqueous deposition	Effusive	Lava flows from land into water.	Congealing flows, pillows, broken pillows. Explosive disruption.	Massive flows. Broken pillow complexes and hyaloclastites. Explosive disruption may produce littoral cones.
	Explosive	Turbulent suspensions in air.	Fallout on water to bottom.	Thin, well sorted, normally graded beds. Sharp bases, bioturbated tops. May be in deep sea 100s of km from source.
		Pyroclastic flows from land into water (may be destroyed by explosive disruption upon entering water).	Dilute suspension fallout from tops of flows.	Thin, well sorted, normally graded. Partly derived from air fall and surge from land which does not enter water. Rest on turbidites. May be reworked by bottom currents.
			Turbulent flows from tops of mass flows.	Thin sequence of fairly well sorted beds, may be cross-bedded. Rest on pyroclastic flows. Tops may be reworked by bottom currents.
			Laminar mass flows.	Thick, poorly sorted, poorly bedded. May be welded to base.
Subaerial	Effusive	Flows	Congealing flows	Massive forms, sometimes mostly rubble. No pillows.
	Explosive	Ballistic; turbulent suspension in air.	Fallout from air	Thin to thick, well sorted; may show normally graded bedding.
		Pyroclastic flows.	Fallout. Derived from tops of flows.	Thin, well sorted beds.
			Turbulent suspensions (pyroclastic surge.)	Thin sequences, fairly well sorted, commonly well bedded, may be cross-bedded.
			Laminar mass flows.	Thick to thin, poorly sorted, massive to poorly bedded, welded to non-welded.

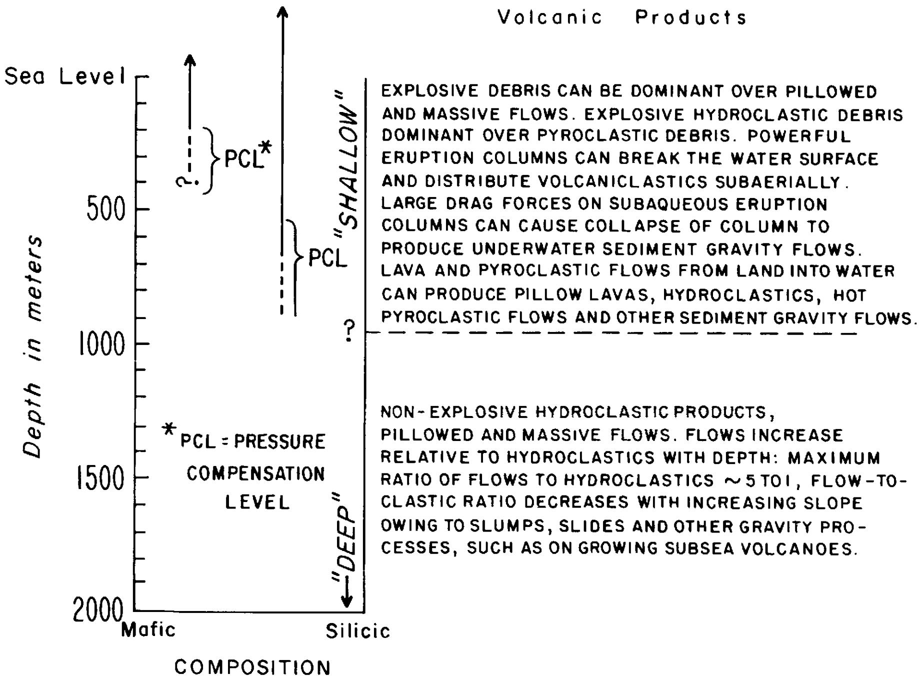

FIG. 1. Submarine volcanic products related to water depth and composition. Pressure compensation levels (PCL) are not specifically known and depend upon magmatic gas pressures, volumes and expansion rates relative to pressure exerted by the water column.

(Fisher & Schmincke 1984). It might exceed 1000 m for silicic and volatile-rich mafic alkalic magmas, but apparently is less than 500 m for most mafic basaltic magmas (McBirney 1963), well within the range of marine shelves that fringe continents and many islands, and of the tops of some submerged islands. Geological evidence suggests that the PCL for explosive alkali basaltic volcanism may be 500–1000 m (Staudigel & Schmincke 1981), but it is less than 200 m below water level for most basaltic eruptions (Moore & Fiske 1969; Jones 1970; Fleet & McKelvey 1978; Allen 1980; Furnes *et al.* 1980).

In addition to pyroclastic fragmentation caused by explosive expansion of volatiles from within magma, there are hydroclastic processes, i.e. fragmentation caused by contact of magma and water (Fisher & Schmincke 1984). Hydroclastic processes can be non-explosive or explosive. Explosive processes can only take place above the PCL, but non-explosive fragmentation can occur at all depths.

Non-explosive hydroclastic processes include granulation and thermal spalling due to stresses set up in magma undergoing rapid quenching (Rittmann 1962; Carlisle 1963; Honnorez 1972), kinetic processes of falling, slumping and breaking of detached pillows on steep slopes, rupturing of growing pillows (Rittmann 1962) and implosions of evacuated pillows where water pressure exceeds internal pillow pressures (Moore 1975).

Explosive hydroclastic processes occur when (i) pore water in rocks is rapidly vaporized by an underlying magma body, or when (ii) water becomes mixed with magma beneath the surface, or under a lava at the surface. Mixing processes are complex (e.g. see Bennett 1972; Colgate & Sigurgeirsson 1973; Wohletz 1980; Kokelaar 1983). Pillowed lava flows and hyaloclastites, either mafic or silicic in composition, are prime evidence of subaqueous environments. In areas where fossils or other evidence (e.g. turbidites, carbonates, and facies associations such as shelf, slope or submarine fan) are absent, pillow lavas are commonly the only evidence of a subaqueous environment.

Mafic pillowed and sheet lava are the two chief forms of lava flows at mid-ocean ridges, as directly observed by submersibles and inferred from drilling through the ocean crust (Bellaiche *et al.* 1974; Needham & Francheteau 1974; Ballard & Moore 1977; Lonsdale 1977; Ballard *et al.* 1979). Mafic volcaniclastic debris is commonly overlooked in deep-sea environments because rock recovery from drill cores averages about 20% and gives a bias in abundance toward massive lava. Robinson *et al.* (1980), however, obtained up to 20% hyaloclastite in Cretaceous crust on the Bermuda Rise from three drill site cores with an average recovery of 70%.

Subaqueous mafic pillow flows and hyaloclastites are most abundant but silicic pillows and hyaloclastites have recently been described (Bevins & Roach 1979; Furnes *et al.* 1980; De Rosen-Spence *et al.* 1980). According to Dimroth *et al.* (1979), in comparison to subaqueous basic lavas, subaqueous silicic flows are thicker and less extensive, lava pods and lobes are similar but larger, quickly chilled rinds on pods and lobes are thicker, and vesicles are larger.

It is generally thought that massive silica-rich lava flow sheets cannot form in submarine environments. However, thick, massive and extensive (up to 2000 km^2) dacite-andesite and rhyodacite lava flows (Cas 1978) are interbedded and conformable with marine Palaeozoic flysch deposits in Australia. Cas (1978) suggests that the lava flows were erupted in water depths great enough to prevent escape of most volatiles. The flows remained mobile because volatiles were retained, and the rate of viscosity increase from cooling was considerably less than the rate of emplacement.

Transport processes and deposits

There are four principal kinds of clastic (volcanic or non-volcanic) deposists on the sea floor. These are, from proximal to distal locations, slump and slide (olistostromes), mass flow, turbidite and pelagic rain (suspension fallout) (Fig. 2). In volcanic as well as in non-volcanic regions, initial debris can be epiclastic, but in volcanic regions there also can be subaqueous lava flow, pyroclastic flow and suspension fallout deposits derived directly from subaqueous or subaerial eruptions (Tables 1 and 2). Moreover, pyroclastic material deposited on steep volcano slopes or shelf margins can be later remobilized as slides and transformed to submarine lahars, turbidity currents and suspension fallout (Fig. 3). The presence of primary and reworked pyroclastic and hydroclastic debris indicates penecontemporaneous volcanism (and tectonism) whereas epiclastic volcanic debris may not.

Flow transformations

Middleton & Hampton (1973, 1976) distinguish between (i) fluid gravity flow in which fluid is moved by gravity and entrains or drives the sediment parallel to the bed (e.g. rivers, ocean currents), and (ii) sediment gravity flow in which movement is by gravity and the sediment motion moves the interstitial fluid (e.g. grain flow, debris flow) (Fig. 4). Lowe (1979) and Nardin *et al.* (1979) have classified sediment gravity flows based upon flow rheology and particle support mechanism.

Sediment gravity flows exhibit the phenomenon of *flow transformation* (Fisher 1983). This term refers to changes within a flow from laminar to turbulent, or vice versa, in turn related chiefly to particle concentration, thickness of the flow and flow velocity (slope). Transformations from slides or slumps to flows (e.g. Naylor 1980) could be considered in another class of transformations but are not discussed here. Stanley *et al.* (1978) use the word 'transforma-

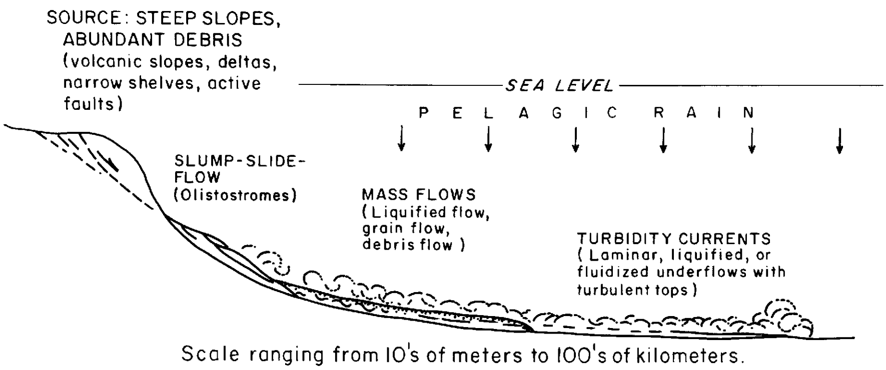

FIG. 2. Conditions for initiation of flows and spectrum of main subaqueous transport processes in non-volcanic regions, but which also occur on subaqueous slopes of volcanoes (see Fig. 3).

TABLE 2. *Simplified list of particles, processes, environments and rocks in subaqueous settings*

Origin of particles	Processes of emplacement[b]	Kinds of deposits	Sites of origin[b]	Sites of deposition[c]
Hydroclastic[a]	Slides	Olistostromes	Volcanic	Beach
Pyroclastic	Laminar flow	Debris flows	eruption	Delta
Epiclastic	Turbulent flow	Turbidites	Volcano	Shelf
	Suspension fallout	Suspension fallout	slopes	Slope
		Pyroclastic flows	Sedimentary	Fan
			aprons	Plain

[a]Can be of explosive or non-explosive origin.
[b]Subaerial or subaqueous.
[c]Sediments can be supplied directly from an eruption or can be remobilized from steep slopes such as delta fronts, continental slopes, etc. Lava flows can occur in proximal environments.

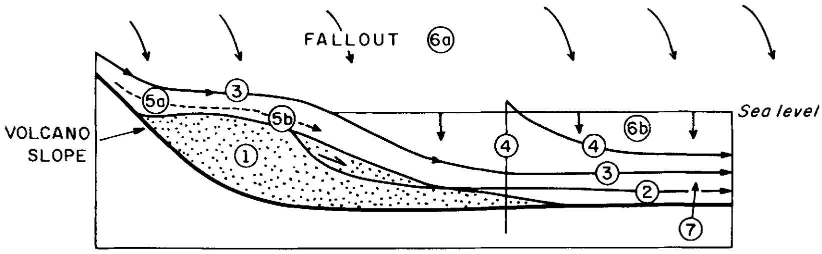

FIG. 3. Sources for subaqueous volcanic debris, and transport processes. This figure combines conditions given in Fig. 2 with volcanics derived directly from an active volcano. (1) Clastic wedge. Hydroclastic, pyroclastic and epiclastic debris. Dominance of particle type determined by type of volcanic activity and periods of inactivity. (2) Slide–debris flow–turbidity current transitions. (3) Pyroclastic flow from land into water. Transitions to subaqueous lahars and turbidity currents. (4) Pyroclastic flows and suspension fallout from powerful shallow underwater eruptions. Column may or may not breach water surface. (5) Lava flows from land into water; (5a) massive flows; (5b) pillows and hyaloclastites. (6) Suspension fallout; (6a) from subaerial eruptions into water; (6b) from underwater eruptions or flow transformations. (7) Deep eruptions below PCL; flows and pillow lavas.

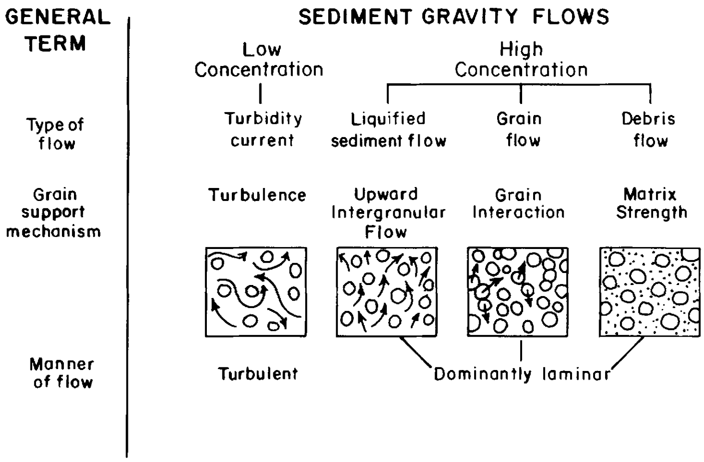

FIG. 4. Classification of sediment gravity flows. Modified from Middleton & Hampton (1973, 1976).

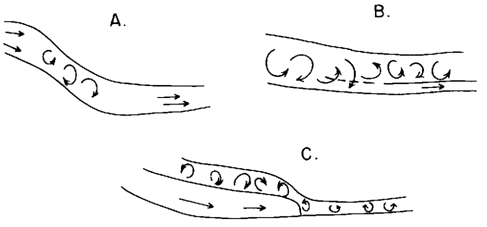

FIG. 5. Schematic representation of (a) body transformations; (b) gravity transformations; and (c) surface transformations. More than one kind of transformation can occur in a single flow and, following separation of flows, transformations can occur again.

tion' but they describe 'transitions' of depositional features, from mass flow to turbidity current deposits, rather than the transformation processes as is done here.

At least three kinds of flow transformations appear to be important in subaqueous environments (Fig. 5):

(i) *Body transformations*, where flow behaviour changes from turbulent to laminar, or vice versa, depending upon the Reynolds and Bingham numbers (see below) within the body of the flow, without significant additions or losses of water.
(ii) *Gravity transformations*, in which flows are initially turbulent but become gravitationally segregated to form a high-concentration laminar underflow with an overriding lower concentration turbulent flow that can move independently.
(iii) *Surface transformations,* where flow separations arise due to the drag effects of water on the top, or nose, of a flow, or due to incorporation of fluid at a hydraulic jump or beneath the nose of a flow.

The concept of flow transformation focuses upon the mass dynamics of flow which chiefly determine sedimentary textures, structures and sequential relationships of massive and laminated units. Various combinations of these three kinds of flow transformations evolving from one another are explained in Fig. 5. Also, fine grained sediment may be thrown into suspension during transformation to produce suspension fallout. *Fluidization transformations* develop by elutriation of fine particles by upward moving fluids from a dense phase bed to produce a turbulent dilute phase above the bed. This type of transformation is best known in gas–solids systems and is believed to produce ash cloud surges from the upper parts of hot pyroclastic flows (Sparks 1976; Fisher 1979; Sheridan 1979; Wilson 1980; Moore & Sisson 1981), and possibly from base surges (Wohletz & Sheridan 1979). Fluidized water-saturated mass flows may occur in subaqueous settings (Middleton & Hampton 1973, 1976; Carter 1975), but the products of elutriation have not been described or perhaps not recognized in the subaqueous realm. As discussed below, in the section on subaqueous pyroclastic flows, gas–solids suspensions might also occur beneath water.

The phenomenon of body transformations is suggested by the experiments and speculations of Kuenen (1952) and Morgenstern (1967) who proposed that slumps or debris flows can change to turbidity currents, with no change in water content, when their velocity is great enough to produce internal turbulence. This can be related to Reynolds and Bingham numbers. Middleton (1970) speculated that changes, from laminar to turbulent and back to laminar flow can occur in large subaqueous grain flows. Fisher's (1971) description of the Parnell Grit, New Zealand, where there are large shale rip-ups in a massive 9 m thick, poorly sorted bed with an inversely graded base, suggests that transformation from turbulent to laminar flow occurred but does not demonstrate a preceding laminar to turbulent transformation.

Gravity transformations are shown experimentally to occur in high-concentration turbidity currents with the development of a 'quick bed' (Middleton 1967). The quick bed is a laminar flowing grain flow and/or fluidized sediment flow (Gonzalez-Bonorino & Middleton 1976) which forms at the base of an initially turbulent high-concentration turbidity current (solids concentration $\geq$ 30% by volume) after passage of the head. Lowe (1982) describes gravity transformations and resulting deposits in high-concentration turbidity currents. He concludes that individual flows become gravitationally segregated, with low-density flows evolving from high-density flows as grain concentration increases toward the base of a flow. Gravity transformations also occur subaerially where initially turbulent pyroclastic flows develop into low-concentration, turbulent pyroclastic surges flowing above high-concentration, laminar block-and-ash flows (Fisher *et al.* 1980; Fisher & Heiken 1982). The phenomenon has also been invoked to explain features of the blast deposits derived from the 18 May 1980 eruption of Mount St Helens (Moore & Sisson 1981). The pyroclastic surge may undergo continued gravity transformation to develop a secondary block-and-ash flow.

Hampton (1972) reviews various subaqueous transformations and shows how surface transformations can occur whereby sediment stripped from the nose and surface of a subaqueous debris flow becomes turbulently mixed with water to form a turbidity current that continues beyond the debris flow. Another kind of surface transformation is described by van Andel & Komar (1969) and Komar (1971). They suggest that at the base of slopes, rapid flows slow down and may go from supercritical conditions (Froude No. > 1) to subcritical (Froude No. < 1). Turbulence develops at this transformation which is known as a hydraulic jump, thereby incorporating ambient fluid into the flow and reducing its density. A further type is described by Allen (1971) who concludes that the frontal parts of slumps may be transformed into turbidity currents by the inflow and mixing of water at their bottom through 'tunnels' that occur along their fronts. The formation of lahars from hot pyroclastic flows that flow into water (e.g. Macdonald 1972) could be classed as a surface transformation.

The manner by which particles are supported in the final stages of flow determines in large part the textures and structures of a bed. The manner of support is a direct function of the particle concentration and the amount of fine grained cohesive sediment mixed with porewater, which in turn greatly affect whether or not laminar to turbulent or turbulent to laminar transformations take place.

The following sections deal mainly with subaqueous debris flows and pyroclastic flows, and their flow transformations. The large and important topic of olistostromes is neglected. They are commonly regarded as bouldery mudstones that originate from submarine slides and slumps (e.g. McBride 1978), although Naylor (1981) cautions that the term should be used descriptively for bouldery mudstones without reference to origin.

Debris flows and lahars

Lahars are debris flows composed of volcanic materials. They may form deposits in subaerial or subaqueous environments, and can originate in several ways (Table 3). Lahars can develop differently from non-volcanic debris flows, but flow and depositional processes are essentially similar. They are sediment gravity flows in which the large fragments (sand, gravel, boulders) are supported mainly by the yield strength of the matrix (Fig. 4). Yield strength in many debris flows is mainly a function of water content and the type and amount of clay which together form the pore fluid (Hampton 1975). Debris flows are modified Bingham plastics (Johnson 1970) in which internal shear stress (τ) must overcome matrix cohesion, a frictional resistance, and internal viscous forces, in order to flow (Middleton & Hampton 1973). Debris flows have a non-flowing central region (rigid plug) where internal shear stress is less than the yield strength, τy. The competence of a debris flow (D) is defined by Hampton (1975) as the diameter of the largest clast that it can carry ($D = 8.8\ \tau y\ g^{-1}$) ($\rho_{\text{clast}}-\rho_{\text{matrix}}$). Much current research on debris flow deals with identifying

TABLE 3. *Origin of subaqueous lahars*

I. Direct and immediate results of eruption
  1. Movement of lahars directly from land into water. Subaerial lahars formed by:
     (a) Eruption through crater lake, snow or ice.
     (b) Heavy rain during eruption.
     (c) Flow of hot pyroclastic material into rivers or onto snow or ice.
  2. Hot pyroclastic flow that moves directly from land into water and becomes mixed with the water.

II. Indirectly related to eruptions
  1. Triggering by earthquakes or sudden liquefaction of water-soaked, pyroclastic or hydroclastic debris previously deposited in shelf environments, or on steep submerged volcano slopes.
  2. Reworking of subaerially deposited pyroclastic debris which enters water from land. Subaerial origin can include:
     (a) Triggering of water-soaked debris by earthquake.
     (b) Bursting and rapid drainage of crater lakes.
     (c) Dewatering of large avalanches originating from collapse of volcano side.

III. Not related to contemporaneous volcano activity
  1. Epiclastic or hydrothermally altered material which becomes water-soaked and moves from land into water.
  2. Epiclastic material on steep coasts and frontal parts of deltas and shelfs which becomes mobilized during slumps and slides.

environments of deposition (especially marine), as aids in constructing facies models and determining tectonic environments (e.g. Stanley *et al.* 1978; Leitch & Cawood 1980; Nemec *et al.* 1980; Swarbrick & Naylor 1980; Gloppen & Steel 1981).

The plastic behaviour of debris flows is the result of high concentration of particles relative to trapped water, and the abundance of fine compared with coarse grained particles. A relatively small percentage (< 10%) of clay-size fragments has a surprisingly large effect on inhibiting turbulence (Hampton 1972). High concentration, irrespective of particle sizes, inhibits turbulence, results in high bulk density and large apparent viscosity values, and contributes to high strength properties. Such flows can support large particles despite laminar flow, and come to rest ('freeze') with steep fronts when internal shear stresses decrease below the threshold strength on low slopes. Moreover, they can flow over soft erodible materials without significantly disturbing the underlying bed. Possibly, they can flow beneath water as far as 500 km, over slopes as low as 0.1° (Embley 1976).

Although debris flows move in laminar fashion (Johnson 1970), turbulence might develop within the body of a flow without mixing with ambient water (a body transformation). The onset of turbulence in Bingham plastics is a function of the ratio of the Reynolds number ($\mathrm{Re} = Ud\rho/\mu$) to the Bingham number ($\mathrm{B} = \tau_c d/\mu U$) where U = average velocity, d = thickness of the flow, ρ = density, μ = dynamic viscosity and τ_c = strength of the plastic. Figure 6 shows that turbulence occurs at $\mathrm{Re} \geqslant 1000$ B. This is equivalent to $\rho U^2/\tau_c \geqslant 1000$ (i.e. $\mathrm{Re/B} \geqslant 1000$), which Hiscott & Middleton (1979) call the Hampton number. Hiscott & Middleton (1979) give an excellent review of the criteria for turbulence, calculations for strength of debris flows, and depositional processes of mass flows (also see Lowe 1976, 1979, 1982). Turbulence, therefore, can develop within the body of a debris flow, despite high concentration, if it is large and velocity is high enough (see also Middleton 1970). If turbulence develops, erosion and mixing with bottom materials and ambient water may occur. This in turn can modify the behaviour of the fluid. Entrainment of clays would tend to reduce turbulence (Hampton 1972), but entrainment of water could lower the relative solids concentration which would favour development of turbulence. With enough water, a transformation from laminar to turbulent flow could be irreversible, although gravity transformations could occur later.

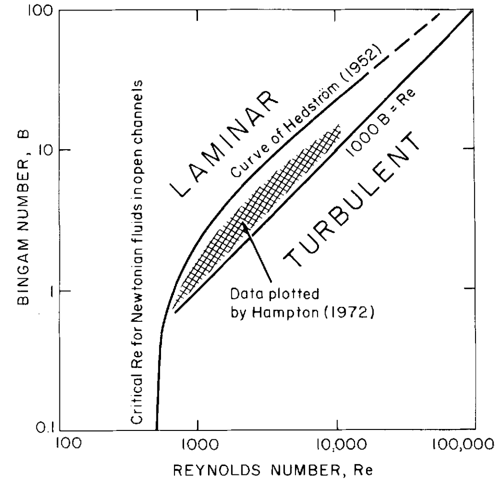

FIG. 6. Relationship of Bingham to Reynolds number for development of turbulence in a Bingham plastic. (After Hiscott & Middleton 1979.)

Plugs are easily deformable masses of debris that form within the interiors of debris flows where shear stresses fall below the strength of the fluid, and within which there is little internal shearing motion. The plug rides on an underflow of mobile debris believed to be laminar (Johnson 1970, p. 500; Hampton 1972). Movement in the underflow occurs because stresses are high enough (a function of velocity and weight). As velocity decreases, stresses decrease, and the lower part of the flow progressively 'freezes' as the yield strength is reached. Thus, the underflow progressively becomes part of the plug. Freezing locks in the textures, structures and fabrics inherited by the plug from upstream before the internal yield strength of the flow is reached.

Characteristic features of deposits from subaerial and subaqueous high-concentration flows include poor or no internal bedding, matrix support of large fragments, poor sorting, and inverse grading in a narrow zone at their base. Inverse grading has been recorded and discussed by many authors. For example in lahars by Schmincke (1967), in subaqueous debris flows and other kinds of sedimentary gravity flows by Fisher & Mattinson (1968), Hiscott & Middleton (1979) and Naylor (1980), and in subaerial pyroclastic flows (in which water or abundant clay is not important in the transport or depositional process) by

Sparks (1976). Elongate particles in these deposits are commonly oriented roughly parallel to the depositional surface or are imbricated. Such orientation is most strongly developed near the base. Basal contacts are commonly non-erosive, and thick, coarse grained debris flows may overlie fine grained, easily erodible material in sharp contact. Large particles may be impressed upon fine grained sediments due to loading. These features occur in subaqueous volcanic debris flows as well as in subaerial counterparts (Fisher 1971), and they are used to recognize debris flows in present-day marine basins (e.g. Carey & Sigurdsson 1980; Sigurdsson *et al.* 1980) and within the geological record (e.g. Cook *et al.* 1972; Cas 1979; Lewis 1976; Hiscott & Middleton 1979; Swarbrick & Naylor 1980; Leitch & Cawood 1980). Subaqueous debris flows appear to differ from subaerial debris flows in several respects, including the ratio of maximum particle size (MPS) to bed thickness, textures, degree of grading and imbrication, and association with other facies (Fig. 7). The bed thickness of subaqueous debris flows is commonly 3–10 times greater than the maximum fragment size, whereas in subaerial deposits, bed thickness is only 2–4 times greater (Nemec *et al.* 1980; Gloppen & Steel 1981). Many subaerial debris flows are abruptly overlain by a thinner, finer grained and better sorted mantle of massive conglomerate or laminated and cross-bedded granule sandstone, interpreted to have been winnowed by water flow following deposition of the debris flow. Commonly the subaqueous deposits are capped by massive or rippled fine grained sandstone which is gradational with the underlying conglomerate. The sandstone is apparently emplaced at the same time as the underlying material rather than the result of later reworking. Subaqueous debris flows can be expected to interfinger with lacustrine or marine sediments.

A B

FIG. 7. Comparison of depositional features of (a) subaerial debris flows, and (b) subaqueous debris flows. (After Nemec *et al.* 1980.)

A subaqueous volcanic debris flow in the Lesser Antilles (Carey & Sigurdsson 1980) is referred to as a sediment gravity flow or pyroclastic gravity flow by Sigurdsson *et al.* (1980), because the structures and textures are similar to subaerial debris flows. The deposit originated from pyroclastic flows which entered the sea and became mixed with water, as indicated by incorporated pelagic clay. The clay content increases with distance from the source, thereby indicating continued mixing during transport. Thus it appears that the flow was turbulent during most of its transport but was transformed (body transformation) to laminar flow in the last stages of movement. There appears to have been little difference between the flow processes of the sediment gravity flow described by Carey & Sigurdsson (1980) and those of the massive division of subaqueous pyroclastic flows described by Fiske & Matsuda (1964) and Bond (1973) (see below).

Large amounts of subaqueous volcaniclastic debris were deposited as debris flows within an Archaean volcaniclastic sequence in the Noranda region, Quebec (Tasse *et al.* 1978). In proximal sections, a significant portion of debris flows (type A beds, Fig. 8) are massive or inversely graded, with rare parallel laminations. Distally, the sequence has a higher percentage of normally graded beds and beds with parallel laminations. The lateral changes can be attributed to surface transformations in developing turbidity currents. Their facies variations in terms of mean bed thicknesses and primary grading features are diagrammatically illustrated in Fig. 8.

Subaqueous pyroclastic flows

Subaqueous pyroclastic flows can develop from pyroclastic flows that move into water from land (e.g. 1902 Mt Pelée: Lacroix 1904). Also, they are postulated to form from entirely underwater eruptions (Fiske 1963; Fiske *et al.* 1963; Fiske & Matsuda 1964; Kokelaar *et al.*, this volume). Many workers do not believe that hot pyroclastic flows can be deposited under water, although some authors report welded tuff from marine sequences (see Table 4). Moreover, Sparks *et al.* (1980b) present theoretical arguments suggesting that welding can occur beneath water. Wright & Mutti (1981) discuss many of the problems encountered in interpreting subaqueous pyroclastic flows.

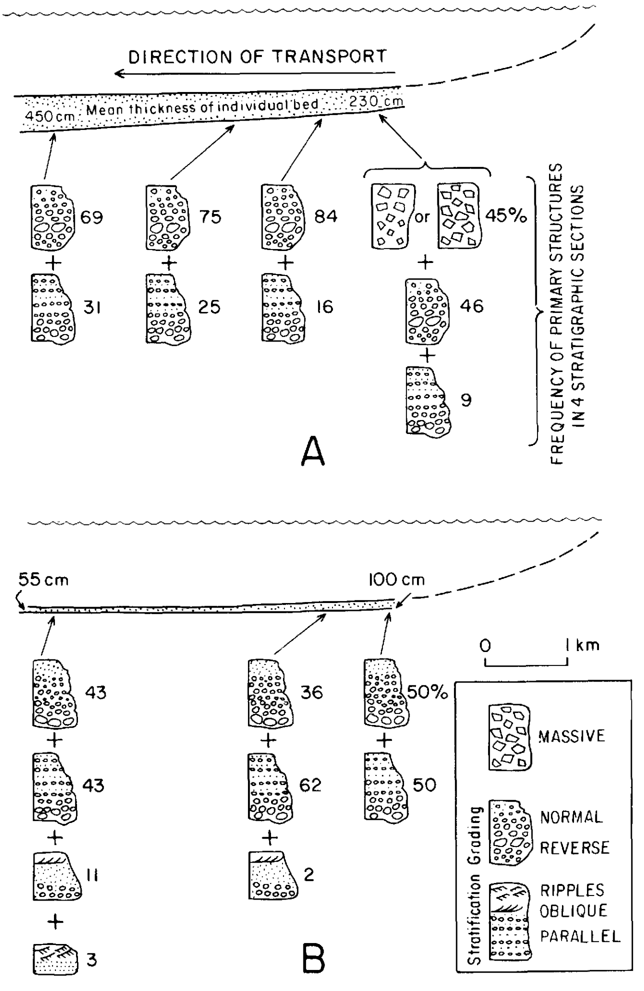

FIG. 8. Lateral variations of primary structures and mean bed thickness in Archaean subaqueous pyroclastic flow deposits, Canada. (a) Mass flow deposits. (b) Turbidite deposits as interpreted by Tasse *et al.* (1978). (From Lajoie 1979.)

Non-welded deposits

Evidence for subaqueous deposition in ancient deposits is determined indirectly on stratigraphical grounds, using sedimentary sequence associations (facies models) or interbedding with pillow lavas or fossiliferous sediments (Fiske & Matsuda 1964; Howells *et al.* 1973; Yamada 1973; Niem 1977). Subaqueous pyroclastic flow deposits consist of lithic fragments, crystals, glass shards and pumice in variable proportions and sizes. Shale rip-ups and blocks up to several metres in diameter derived from underlying beds may be incorporated near the base of a sequence (Yamada 1973; Niem 1977).

Shattered crystals are common in subaqueous pyroclastic flows (Fiske & Matsuda 1964; Fernandez 1969; Yamada 1973; Niem 1977; Fig. 9), as are glassy fragments produced by sudden quenching. Fine perlitic cracks are characteristic of essential, non-vesicular, vitric clasts (Kato *et al.* 1971).

Non-welded subaqueous pyroclastic flow deposits characteristically are massive to poorly bedded and poorly sorted in their lower part (lower division), and are thinly bedded in their upper part (upper division) (Fig. 10). The lower division can form 50% or more of the total sequence (Fiske & Matsuda 1964; Bond 1973; Niem 1977). The two-division sequence is interpreted in terms of a waning, initially voluminous underwater eruption (Fiske & Mat-

TABLE 4. *Subaqueous pyroclastic flow occurrences (from Fisher & Schmincke 1984)*

Environment	Age	Location	Welding and temperature	Remarks	Reference
Marine	1883	Krakatau, Java	Non-welded.	Generated destructive tsunamis.	Self & Rampino 1981
Marine	28,000 yr BP	Grenada Basin, Lesser Antilles	Non-welded. Resemble debris flow deposits.	Derived from subaerial hot pyroclastic flows on Dominica.	Carey & Sigurdsson 1980; Sigurdsson *et al.* 1980
Marine	Late Quaternary	Subaqueous west flank, Dominica, Lesser Antilles	Deposits not cored.	Deposits traceable to subaerial flows (block and ash; welded ignimbrite).	Sparks *et al.* 1980a
Marine	Middle Miocene	Santa Cruz Island, Calif., U.S.A.	Non-welded. Pumice-rich. Massive beds, some with pinkish oxidized tops.	Within a marine sequence.	Fisher & Charleton 1976
Marine	Miocene	Tokiwa Formation, Japan	Non-welded. Pumice-rich.	Interbedded conformably with fossil-rich beds.	Fiske & Matsuda 1964
Marine	Oligocene	Is. of Rhodes, Greece	Non-welded.	Deep water, marine.	Wright & Mutti 1981
Marine	Oligocene–Eocene	Mt Rainier Natl Park, Wash., U.S.A.	Non-welded. Massive lower to bedded upper divisions.	Marine shelf deposition inferred.	Fiske 1963; Fiske *et al.* 1963
Marine	Early Oligocene–Late Eocene	Philippines	Non-welded. Pyroclastic turbidites.	Interbedded with marine limestone.	Garrison *et al.* 1979
Marine	Paleogene to Cretaceous	Philippines	Non-welded to welded (?) tuff.	Within a marine sequence.	Fernandez 1969
Marine	Lower Mesozoic	Sierra Nevada, California, U.S.A.	Metamorphosed (shards not preserved).	Ponded in shallow marine calderas.	Busby-Spera 1981a,b
Marine	Permian	Alaska, U.S.A.	Non-welded. Massive lower to bedded upper divisions.	Structures and sequence similar to Tokiwa Formation, Japan.	Bond 1973
Marine	Mississippian	Oklahoma and Arkansas, U.S.A.	Non-welded. Turbidite structures.	Interbedded within marine flysch sequence.	Niem 1977
Marine	Ordovician	Ireland	Welded.	Associated with shallow-water deltaic sediments.	Stanton 1960; Dewey 1963
Marine	Ordovician	Wales, United Kingdom	Welded.	In marine sequence. Traceable to subaerial pyroclastic flows.	Francis & Howells 1973; Howells *et al.* 1979; Howells & Leveridge 1980
Marine	Lower Ordovician	S. Wales, U.K.	Welded.	Within marine sequence.	Lowman & Bloxam 1981
Marine	Archaean	Rouyn-Noranda, Canada	Non-welded.	Structures and sequence similar to Tokiwa Formation, Japan.	Dimroth & Demarcke 1978; Tasse *et al.* 1978
Lacustrine	Plio–Pleistocene	Japan	Non-welded. Resemble turbidites.	Interbedded with lacustrine rocks. Flow from land into water.	Yamada 1973
Lacustrine	Mio–Pliocene	Japan	Non-welded. Thermoremanent magnetism suggests emplacement at 500°C	Massive units resembling debris flows.	Kato *et al.* 1971; Yamazaki *et al.* 1973

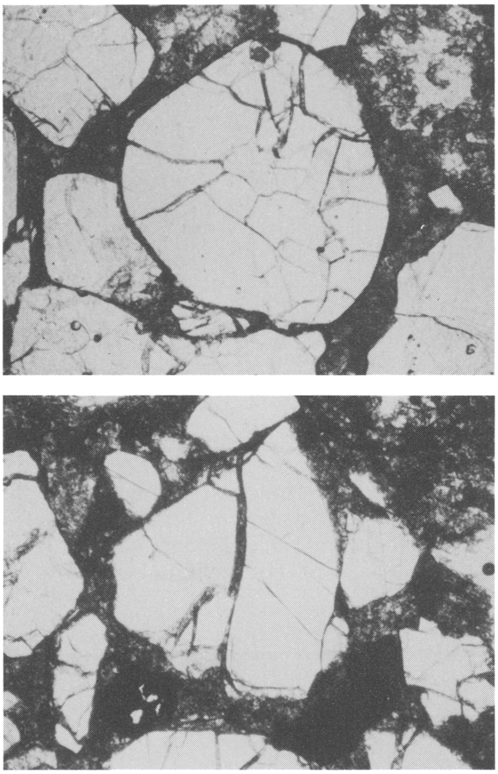

FIG. 9. Shattered quartz crystals from a subaqueous pyroclastic flow deposit, Eugenia Formation, Baja, California (J. Hickey, pers. comm. 1982).

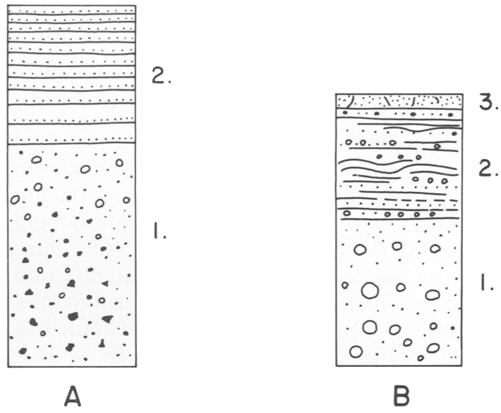

FIG. 10. Examples of non-welded subaqueous pyroclastic flow deposits. (a) Example interpreted to have developed from waning eruption by Fiske & Matsuda (1964); massive division (1) overlain by doubly graded sequence (2). (b) Obispo Tuff, interpreted to have developed from passage of one flow or several pulses of flow (Fisher 1977). Massive division (1), overlain by (2) sequence of planar to wavy beds or beds with low-angle cross-beds, fine to medium grained tuff to lapilli tuff layers, and (3) capped by a thin, very fine grained bioturbated tuff of probable suspension fallout origin. Interval (2) may reflect several pulses of flow, possibly representing retrogressive slumping (Morgenstern 1967). Pumice lapilli commonly decrease in size from bottom to top indicating that the tuff was water logged prior to flow, thereby supporting the idea that flow was remobilized pyroclastic debris from an unstable subaqueous slope.

suda 1964). Yamada (1973, this volume), however, recognizes five divisions similar to a Bouma sequence (Bouma 1962). A 2-fold sequence capped by a thin, very fine grained, bioturbated pelagic layer occurs in the Obispo Tuff, California (Fig. 10b), where basal contacts are generally sharp on undisturbed strata, although flute marks, grooves and load casts occur locally. Where bottom contacts are plane and sharp, missing underlying strata may indicate erosion (Yamada 1973).

The lower division of a subaqueous pyroclastic flow unit is a single, coarse grained bed commonly lacking internal structures or laminations (Fiske & Matsuda 1964; Bond 1973), although some beds have incipient cross-laminations outlined by a crude orientation of coarse fragments (Kato *et al*. 1971; Yamazaki *et al*. 1973). Larger and more dense fragments generally occur near the base of the massive divisions, with crystals, glass shards and pumice becoming more abundant toward the top. Dense, lithic fragments may be inversely graded near the base, as in debris flows (Yamada 1973).

The upper division of non-welded subaqueous pyroclastic flow deposits is commonly composed of many thin, fine to coarse grained ash beds. In some sequences, each bed may be normally graded, and the entire sequence of beds becomes finer grained upward. This is called a doubly graded sequence (Fiske & Matsuda 1964). Not all upper divisions are doubly graded (Niem 1977).

Double grading is interpreted by Fiske & Matsuda (1964) to signify contemporaneous waning subaqueous volcanism with deposition from thin turbidity flows following deposition of the massive bed. Deposits without double grading, such as those described by Niem (1977) (see also Fig. 10b), are taken to signify sloughing from oversteepened pyroclastic slopes on the edge of a volcano where transformations from slides to sediment gravity flows could occur.

Welded deposits

Most controversial are welded pyroclastic flow deposits interpreted to be subaqueous on stratigraphic grounds. The best known example is from the Ordovician of Snowdonia in Wales. There, a welded tuff within the Capel Curig Formation is interpreted to be the submarine correlative of subaerial welded tuff (Francis & Howells 1973; Howells *et al.* 1973; Howells *et al.* 1979; Kokelaar *et al.*, this volume a,b). Francis & Howells (1973) suggest that subaerially produced pyroclastic flows entered the sea and remained hot enough to weld after subaqueous emplacement. The subaerial welded tuff consists of an uninterrupted sequence of two or more flow units of greater combined thickness (70 m) than correlative subaqueous units (< 50 m). The base of the subaerial tuff is planar, contains fiamme and overlies a conglomerate with a reddened top, and welding is restricted to the middle of the sheet. The subaqueous pyroclastic flow deposits are welded to their base, and are interbedded with marine siltstones containing brachiopods. Basal contacts of the subaqueous welded tuffs tend to be irregular with zones resembling large-scale load and flame structures. These flow deposits are massive in the lower and central portions and grade imperceptibly upward through ill-defined, faintly bedded zones to evenly bedded tuff which locally has broad, low-angle cross-bedding suggestive of marine deposition. Their tops are in sharp contact with overlying sandstone with similar broad cross-bedded structures. Kokelaar (1982) accounts for the large load-structure protrusions at the base of the subaqueous welded tuffs by fluidization of subjacent wet sediments following emplacement of a hot pyroclastic flow.

Origin of subaqueous pyroclastic flows

Three principle ways in which subaqueous pyroclastic flows can be generated are given in Fig. 11. Massive lower divisions with finer grained bedded tops can develop in each situation. Mixing with water by flow transformation to form sediment gravity flows and turbidity currents can also occur in each case. The occurrence of subaqueous welded tuff in Wales correlative with subaerial welded tuff (Francis & Howells 1973) suggests that hot pyroclastic flows that enter water from land (case A) may do so without mixing with water and retain enough heat to weld. As shown by Sparks *et al.* (1980b), welding is theoretically possible. This would be likely to occur where changes in slope are insufficient to cause laminar to turbulent body transformations. Such transformations are more likely to occur where sudden steepening of slope could cause turbulence and mixing, thereby producing a cool sediment gravity flow (Sigurdsson *et al.* 1980). Also as shown for case A (Fig. 11), turbulent ash cloud surges, which can develop above dense flows on land, can flow over the water (Anderson & Flett 1903). Additionally, surface transformations on the underwater flow could produce turbidity currents. In each instance, thin, graded beds with small total volume can develop by fallout at the top of flow units.

For case B (Fig. 11), Fiske & Matsuda (1964) envisage an eruption column emerging with great force from an underwater vent. Large amounts of material are ejected into water, fall back to the sea floor, and form a

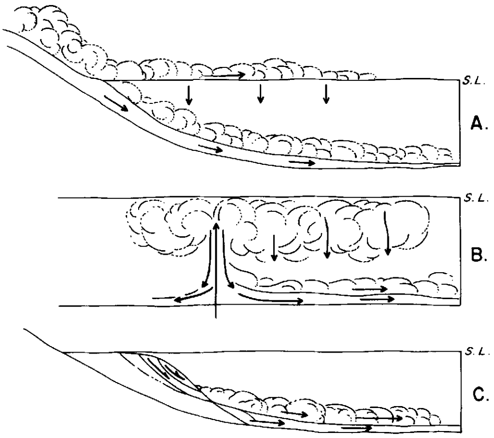

FIG. 11. Means by which subaqueous pyroclastic flows can develop. (a) Hot pyroclastic flow from land into water (Francis & Howells 1973; Yamada 1973; Sigurdsson *et al.* 1980). Density separation of flow and ash cloud surge at air–water interface; surface transformation to form thin turbidite sequence. Flow may weld if not mixed with water, or be transformed to debris flow if mixed. (b) Pyroclastic flow develops from column collapse (Fiske 1963; Fiske & Matsuda 1964); waning eruption may produce overlying laminated doubly graded sequence. Suspended material can form thin fallout layers and floating pumice can form pumice rafts. (c) Sediment gravity pyroclastic flows develop from slumping of unstable slopes composed of pyroclastic debris.

dense water-rich debris flow deposited as a thick, massive layer. In the closing stages of eruption, small turbidity currents entrain pumice lapilli and crystals of plagioclase and quartz which are deposited as thinly graded beds above the massive beds.

Ash from a submarine eruption cloud that breaks the water surface, or that has spread widely away from the vent within the water body, continues to settle after the eruption ends, but in decreasing amounts and becoming finer grained. Thus, there is an insufficient supply to maintain the continuous flow that deposited the thick structureless part of the pyroclastic flow. Instead, intermittent thin turbidity currents follow one another in close succession. As the particle-rich column continues to settle, successively smaller amounts of finer grained and less-dense ash fall out, and the density currents become less frequent and finally end. Thus, the total volume of the upper division is considerably greater than that developed at the top of a single flow entering water from land (case A), although distally Bouma sequences may dominate in case A (Yamada 1973, this volume).

A vertical eruption column formed entirely beneath water, such as that proposed by Fiske & Matsuda (1964), may not develop into hot pyroclastic flows upon collapse because turbulent mixing with water along its margin would be enhanced by drag effects in water which would cool the materials. Possibly, however, a voluminous 'boiling-over' type of eruption without a vertical column (Fisher & Schmincke 1984) could be extruded at rates great enough to produce a flow protected from the water by a carapace of steam and retain enough heat to become welded. As yet, however, no subaqueous vents for welded tuffs have been positively identfied, although one has been postulated by Kokelaar *et al.* (this volume a, b) to explain a welded tuff sequence on Ramsey Island in SW Wales.

Slumping of unstable material (case C) composed exclusively of pyroclastic material is probably common on submerged slopes of volcanoes, but has rarely been documented. It is to be expected, however, that flow transformations from subaqueous lahars to turbidity currents can develop sequences similar to those produced directly from volcanic eruptions. A thick (*c.* 15 m) coarse grained pyroclastic flow within the Obispo Tuff, California is interpreted to have formed by cold water mobilization of pyroclastic material off steep slopes (Fisher 1977).

Subaqueous pyroclastic flows interbedded with lacustrine sediments of Onikobe Caldera, Japan (Yamada 1973), closely resemble turbidite sequences. The flows may have originated from subaerial eruptions at the margin of the caldera, but there is little evidence to indicate whether hot pyroclastic flows entered the water from land, or if slumping of unstable slopes on the margin of the lake initiated the flows.

Sparks *et al.* (1980b) have proposed on theoretical grounds that if a hot pyroclastic flow manages to pass through the air–water interface, subaqueous welding is not only possible, but is enhanced. Hydrostatic pressure increases by 1 bar for every 10 m water depth, and if water is 'drawn' into the hot pyroclastic flow it can be absorbed into glass shards. At increased pressures, water content of glass can be higher, resulting in lower viscosities which favour welding at lower temperatures (Smith 1960; Riehle 1973). Calculations of the viscosity of glass shards as a function of water depth suggest that the viscosity of glass decreases substantially with depth despite cooling of the flow (Sparks *et al.* 1980b). For subaerial pyroclastic flows at 800°C, the viscosity of rhyolitic glass is about 2×10^{11} poise, but, at 755°C under 500 m of water, this is reduced by dissolution of water to an estimated 2×10^{9} poise. Thus glass under the latter conditions would compact much faster than in a subaerial flow under the same uniaxial stress. A major difficulty with the theory, however, is the slow diffusion rate of water into glass. In addition, there is a question of how water could be drawn into a flow and into contact with glass participles. Turbulent mixing is the dominant process by which water can gain access to a flow, but this would also tend to cool or explosively destroy the flow (Walker 1979). A rapid increase of pressure, from loading by the water column, would inhibit volatile escape and is a more likely cause of welding. Water surrounding the hot flows would create a thin vapour carapace around the flow during cooling (Francis & Howells 1973; Yamazaki *et al.* 1973), provided that turbulent mixing does not occur at the surface of the flow.

Submarine ash layers

The initial depositional process of marine fallout ash is by the settling of particles introduced into water from the air or from underwater flows or eruptions. The process therefore depends upon the size, shape and density of fragments, and the vertical and lateral distribution is influenced by outside factors such as wind and water current velocities and directions. For more complete reviews of this subject, the reader is referred to Kennett (1981) and Fisher & Schmincke (1984). Data from widespread submarine fallout layers are useful in many ways. For example, (i) in correlating over long distances between sequences deposited in different environments (Ninkovich & Heezen 1965, 1967; Huang *et al.* 1973; Watkins *et al.* 1978; Keller *et al.* 1978; Hahn *et al.* 1979; Thunell *et al.* 1979; Ninkovich 1979), (ii) in solution of problems of magma genesis and evolution (Scheidegger *et al.* 1978, 1980), (iii) in assessments of volcanic cycles (Hein *et al.* 1978), (iv) in derminations of production rates of tephra and eruption column heights (Shaw *et al.* 1974; Huang *et al.* 1973, 1975, 1979; Watkins & Huang 1977; Ninkovich *et al.* 1978), (v) in estimations of the duration of eruptions (Ledbetter & Sparks 1979), (vi) as aids in determining rates of plate movements and arc polarity (Ninkovich & Donn 1976; Sigurdsson *et al.* 1980), (vii) in determination of prevailing wind and water current directions (Eaton 1964; Nayudu 1964; Ninkovich & Shackleton 1975; Hampton *et al.* 1979), (viii) in making inferences about climatic changes (Kennett & Thunell 1975, 1977) and (ix) in solving problems concerning diagenesis (Coombs 1954; Fisher & Schmincke 1984).

Widespread marine fallout layers can occur in back-arc and fore-arc regions, depending mainly on wind directions, and occur in all marine sedimentary environments. In the Lesser Antilles, marine fallout layers are the dominant volcaniclastic type in front of the arc (Sigurdsson *et al.* 1980), but in the Cordilleran region of western North America they are most abundant behind the Cascades volcanic arc, presently as well as in ancient deposits (Eaton 1964; Dickinson 1974).

Island volcanoes

Irrespective of magma composition and marginal or intra-plate setting, volcanoes in the sea can be considered to evolve through two overlapping stages (Fig. 12). Stage A develops beneath the PCL and stage B above the PCL (L. Ayres, personal communication). As suggested by previous discussions, volcanic island foundations (below the PCL) can be composed of massive and pillowed lava flows with lesser amounts of hyaloclastites (up to 20%). As volcanoes grow and slopes steepen in stage A, talus development and slides cause mechanical fragmentation and movement of debris into deeper water, resulting in a decrease in the lava-to-clastic ratio outward from the source. This is observed on sea-mounts (Stanley & Taylor 1977; Lonsdale & Spiess 1979). Above the PCL, however, there is a wider range of transport processes, kinds of fragments and consequent lateral facies changes in rock types.

Ayres (pers. comm.) infers a shield shape for the foundations of island volcanoes in stage A, with steeper slopes developing in stage B. In Fig. 12, lava-to-clastic ratio is estimated at up to 5:1 in early growth stages and in flatter summit regions of cones, decreasing perhaps to 2:1 or less away from the source as slopes steepen, to entirely clastic on surrounding abyssal plains. In addition to lava flows, the intrusion of sills may be important in the growth of an island volcano. On La Palma, Canary Islands, for example, Staudigal & Schmincke (1981) report 2000 m of sills which contributed significantly to the elevation of the volcano.

Rocks of stage B_1, above the PCL but below water level, can be dominantly explosive hydroclastic debris. At DSDP Site 253 (Ninetyeast Ridge, Indian Ocean) for example, there is at least 388 m of bedded Middle Eocene basaltic

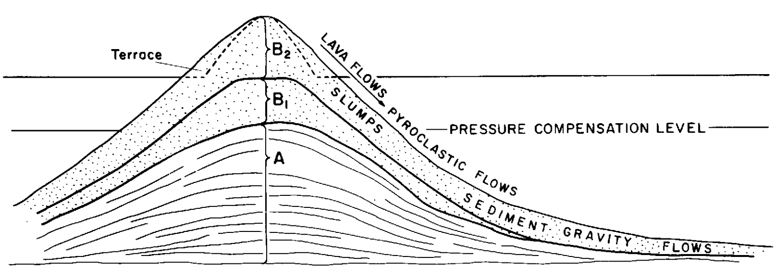

FIG. 12. Generalized model of an island volcano. Stage A, foundation of island volcano beneath pressure compensation level (PCL); lava (L) > hydroclastics (H); may contain a significant volume of sills. Stage B_1, beneath sealevel and above PCL; L < H. Stage B_2, subaerial part; pyroclastics (P) > H > L if andesitic, but L > P if basaltic shield. Clastic apron enlarges rapidly in stage B_2 and sediment gravity flows and fallout deposits increase in flanks and surrounding basin.

hydroclastic debris, believed to have been erupted in about 150 m of water (Fleet & McKelvey 1978).

In stage B_2, lava flows of basaltic islands, such as Hawaii, cap the hydroclastic material as the vents become subaerial (Moore & Fiske 1969). Along the fringes of such islands, explosion debris, pillows and pillow breccias can be generated where lavas flow into the sea (Moore 1975; Peterson 1976; Fisher 1968). Erosion commences to produce epiclastic volaniclastic debris which increases in volume as the volcano grows in size. However, at volcanoes where pyroclastic material is abundant relative to lava flows, such as andesitic stratocones, emerged slopes in stage B_2 are steeper than shield volcanoes, and pyroclastic debris is delivered to the sea as lahars, pyroclastic flows and fallout ash directly from eruptions, or by remobilization of loose pyroclastic deposits. The volumes of pyroclastic material stored in clastic wedges on submerged slopes of andesite volcanoes is consequently greater than for shield volcanoes, and so is the volume of debris deliverered to adjacent basins as sediment gravity flows and suspension fallout. The rate and volume of debris supplied during periods of volcanic activity and inactivity is discussed by Kuenzi *et al.* (1979) and Vessel & Davies (1981). The rate and volume of supply from fringing terraces to adjacent slopes and basins can also be modified by eustatic sealevel changes. During high stands of sealevel, more sediment can be stored in the clastic wedges and less reaches the basins than during low stands, when sedimentation rate and volume of sediment increases in the basins (Klein *et al.* 1979). Primary and reworked volcanic products become more diverse as an island emerges, because erosion is contemporaneous with eruptions, and sediment aprons with unstable slopes can grow more rapidly than before emergence. Reefs can form, and in areas of long-continued volcanism and erosion, fringing shelves become the sites for development of deltas and fan-deltas with deep-sea fans and abyssal turbidite plains farther away (Londsdale 1975).

Facies of volcanic rocks reflect distance to source. Dykes, sills, plutons, lava flows, coarse grained pyroclastic rocks and their reworked equivalents, occur at or near the source. Farther away, lava flows and intrusive rocks disappear and volcaniclastic sediments become finer grained (Dickinson 1968; Mitchell 1970). Fall-out tuffs occur within both of these near- and intermediate-distance regions, but may be dispersed globally and occur at great distances from source volcanoes. These relationships apply to both subaerial and submarine environments.

It is also true that (i) volcanic rocks are emplaced in the order in which they are extruded, and (ii) the eroded products of an inactive volcano tend to be emplaced in reverse order of extrusion. Erosion can extend down to the source plutons (Dickinson & Rich 1972; Ingersoll 1978; Mansfield 1979). This simplified scheme (Fig. 13) is only a basic framework, however, because erosion of a volcanic edifice can alternate with periods of active growth, adjacent volcanoes can supply material to basins at different times, distant volcanoes can supply ash during any part of the constructive or destructive cycles, constant settling of pelagic and hemipelagic material from marine waters can occur, and sediments from land masses and islands can be supplied from opposite sides of a basin. Further complications arise, for example, where moving plates carry source areas away from the depositional site (diverging plates), toward the depositional site (converging plates), or laterally along transform boundaries.

Irrespective of lateral facies changes and despite the many complexities, it is possible to

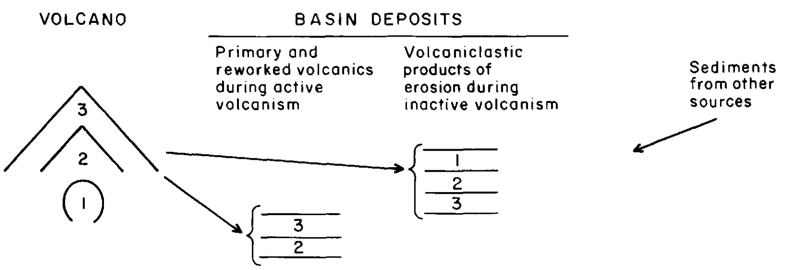

FIG. 13. Simplified stratigraphic scheme for deposition of volcaniclasitc debris from active and inactive volcanoes in adjacent basins. (1) plutonic sources and epiclastic products, (2) and (3) volcanic products in sequence of emplacement from active volcanoes. Erosion of the volcanic pile produces epiclastic debris in inverse order.

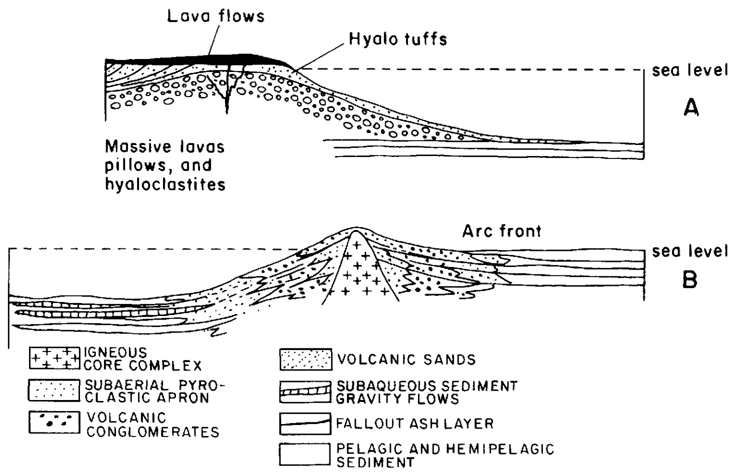

FIG. 14. Facies model of island volcanoes. (a) Newly emergent intra-plate basaltic shield volcano (from Moore & Fiske 1969). (b) Island-arc andesitic stratocone (from Sigurdsson *et al.* 1980).

gain considerable information, about many of the aspects discussed above, from vertical sequences (in outcrop or drill-hole samples), if the origin of clasts can be determined (e.g. pyroclastic, hydroclastic, epiclastic) and depositional processes can be interpreted from rock textures and structures (see Schmincke & von Rad 1979).

Work in the Lesser Antilles has led to a facies model of a calc-alkaline island-arc volcano (Fig. 14) (Carey & Sigurdsson 1978; Sigurdsson *et al.* 1980; Sparks *et al.* 1980a; see also Carey & Sigurdsson, this volume). In the Lesser Antilles, dominant kinds of sediment transport processes and deposit types are highly asymmetric. Volcaniclastic sediments form 34% of the total cored sediment behind the arc and 4% of the total in front of the arc, a proportion which does not significantly vary within 300 km of the arc. Sediment gravity flows of various kinds form 98% of the volcaniclastic sediments in the back-arc basin, and ash-fall layers form 60% of the volcaniclastic sediments in the fore-arc region. This asymmetry is attributed to wind directions, arc slopes (steeper toward the back-arc) and ocean current directions.

Research in the Lesser Antilles documents sedimentary asymmetry associated with arc polarity. Sedimentary asymmetry, however, depends upon arc geometry (facing direction, trend direction of the arc) relative to wind and ocean current directions. Therefore, volcaniclastic facies as related to arc polarity will differ in different arcs.

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References

ALLEN, C. C. 1980. Icelandic subglacial volcanism: thermal and physical studies *J. Geol.* **88**, 108–17.

ALLEN, J. R. L. 1971. Mixing at turbidity current heads, and its geological implications. *J. sediment. Petrol.* **41**, 97–113.

ANDERSON, T. & FLETT, J. S. 1903. Report on the eruption of the Soufriere in St Vincent in 1902 and on a visit to Montagne Pelee in Martinique, Part I. *Philos. Trans. R. Soc. London.* **A200**, 353–553.

BALLARD, R. D. & MOORE, J. G. 1977. *Photographic Atlas of the Mid-Atlantic Ridge.* Springer, New York.

——, HOLCOMBE, R. T. & VAN ANDEL, T. H. 1979. The Galapagos Rift at 86°W. III. Sheet flows, collapse pits, and lava lakes of the rift valley. *J. geophys. Res.* **84**, 5407–22.

BELLAICHE, G., CHEMINEE, J. L., FRANCHETEAU, J., HEKINIAN, R., LEPICHON, X., NEEDHAM, H. D. & BALLARD, R. D. 1974. Rift Valley's inner floor: First submersible study. *Nature, Lond.* **250**, 558–60.

BENNETT, F. D. 1972. Shallow submarine volcanism. *J. geophys. Res.* **77**, 5755–60.

BEVINS, R. E. & ROACH, R. A. 1979. Pillow lava and isolated pillow breccia of rhyodacitic composition from the Fishguard Volcanic Group, Lower Ordovician, SW Wales, United Kingdom. *J. Geol.* **87**, 193–201.

BOND, G. C. 1973. A late Paleozoic volcanic arc in the eastern Alaska Range, Alaska. *J. Geol.* **81**, 557–75.

BOUMA, A. H. 1962. *Sedimentology of Some Flysch Deposits; A Graphic Approach to Facies Interpretation.* Elsevier, Amsterdam, Netherlands.

BUSBY-SPERA, C. 1981a. Silicic ash-flow tuffs interbedded with submarine andesitic and sedimentary rocks in lower Mesozoic roof pendants, Sierra Nevada, California (Abst.). *Bull. geol. Soc. Am.* **13**, 47.

——, 1981b. Early Mesozoic submarine epicontinental calderas in the southern Sierra Nevada, California. *Eos*, **62**, 1061.

CAREY, S. N. & SIGURDSSON, H. 1978. Deep-sea evidence for distribution of tephra from the mixed magma eruption of the Soufriere on St. Vincent, 1902: ash turbidites and air fall. *Geology*, **6**, 271–4.

—— & SIGURDSSON, H. 1980. The Roseau Ash: deep-sea tephra deposits from a major eruption on Dominica, Lesser Antilles Arc. *J. Volcan. geoth. Res. Amsterdam*, **7**, 67–86.

CARLISLE, D. 1983. Pillow breccias and their aquagene tuffs, Quadra Island, British Columbia. *J. Geol.* **71**, 48–71.

CARTER, R. M. 1975. A discussion and classification of subaqueous mass-transport with particular application to grain-flow, slurry-flow, and fluxoturbidites. *Earth Sci. Rev.* **11**, 145–77.

CAS, R. 1978. Silicic lavas in Paleozoic flyschlike deposits in New South Wales, Australia: Behaviour of deep subaqueous silicic flows. *Bull. geol. Soc. Am.* **89**, 1708–14.

—— 1979. Mass-flow arenites from a Paleozoic interarc basin, New South Wales, Australia: mode and environment of emplacement. *J. sediment. Petrol.* **49**, 29–44.

COLGATE, S. A. & SIGURGEIRSSON, T. 1973. Dynamic mixing of water and lava. *Nature, Lond.* **244**, 552–5.

COOK, H. E., MCDANIEL, P. N., MOUNTJOY, E. W. & PRAY, L. C. 1972. Allochthonous carbonate debris flows at Devonian bank ('reef') margins, Alberta, Canada. *Bull. Can. Pet. Geol.* **20**, 439–97.

COOMBS, D. S. 1954. The nature and alteration of some Triassic sediments from Southland, New Zealand. *Trans. R. Soc. N.Z.* **82**, 65–109.

DE ROSEN-SPENCE, A. F., PROVOST, G., DIMROTH, E., GOCHNAUER, K. & OWEN, V. 1980. Archean subaqueous felsic flows, Rouyn-Noranda, Quebec, Canada, and their Quaternary equivalents. *Precambrian Res.* **12**, 43–77.

DEWEY, J. F. 1963. The Lower Paleozoic stratigraphy of central Murrisk County, Mayo, Ireland, and the evolution of the South Mayo Trough. *J. geol. Soc. London*, **119**, 313–43.

DICKINSON, W. R. 1968. Sedimentation of volcaniclastic strata of the Pliocene Koroimavua Group in northwest Viti Levu, Fiji. *Am. J. Sci.* **266**, 440–53.

—— 1974. Sedimentation within and beside ancient and modern magmatic arcs. *In*: DOTT JR, R. H. & SHAVER, R. H. (eds) *Modern and Ancient Geosynclinal Sedimentation.* Spec. Pub. Soc. Econ. Paleont. Miner. **19**, 230–9.

—— & RICH, E. L. 1972. Petrographic intervals and petrofacies in the Great Valley Sequence, Sacramento Valley, California. *Bull. Geol. Soc. Am.* **83**, 3007–24.

DIMROTH, E. & DEMARCKE, J. 1978. Petrography and mechanism of eruption of the Archean Dalembert tuff, Rouyn-Noranda, Quebec, Canada. *Can. J. Earth Sci.* **15**, 1712–23.

——, COUSINEAU, P., LEDUC, M., SANSCHAGRIN, Y. & PROVOST, G. 1979. Flow mechanisms of Archean subaqueous basalt and rhyolite flows. *In*: *Current Res. Part A, Geol Surv. Canada, Paper* **79-1A**, 207–11.

EATON, G. P. 1964. Windborne volcanic ash: A possible index to polar wandering, *J. Geol.* **72**, 1–35.

EMBLEY, R. W. 1976. New evidence for occurrence of debris flow deposits in the deep sea. *Geology*, **4**, 371–4.

FERNANDEZ, H. E. 1969. Notes on the submarine ash flow tuff in Siargao Island, Surigao del Norte (Philippines). *Philipp. Geol.* **23**, 29–36.

FISHER, R. V. 1968. Puu Hou littoral cones, Hawaii. *Geol. Rdsch.* **57**, 837–64.

—— 1971. Features of coarse-grained, high-concentration fluids and their deposits. *J. sediment. Petrol.* **41**, 916–27.

—— 1977. Geologic guide to subaqueous volcanic rocks in the Nipomo, Pismo Beach and Avila Beach areas. *Penrose Conf. geol. Soc. Am. The Geology of Subaqueous Volcanic Rocks.* 1–29.

—— 1979. Models for pyroclastic surges and pyroclastic flows. *J. Volcan. geoth. Res. Amsterdam*, **6**, 305–18.

—— 1983. Flow transformations in sediment gravity flows. *Geology*, **11**, 273–4.

—— & CHARLETON, D. W. 1976. Mid-Miocene Blanca Formation, Santa Cruz Island, California. *In*: HOWELL, D. G. (ed.) *Aspects of the Geologic History of the California Continental Borderland.* Misc. Pub. Pacific Section Am. Assoc. Petrol. Geol. **24**, 228–40.

—— & HEIKEN, G. H. 1982. Mt Pelée, Martinique: May 8 and 20 pyroclastic flows and surges. *J. Volcan. geoth. Res. Amsterdam* **13**, 339–71.

—— & MATTINSON, J. M. 1968. Wheeler Gorge turbidite—conglomerate series; inverse grading. *J. sediment. Petrol.* **38**, 1013–23.

—— & SCHMINCKE, H.-U. 1984. *Pyroclastic Rocks.* Springer, Heidelberg.

——, SMITH, A. L. & ROOBOL, M. J. 1980. Destruction of St Pierre, Martinique by ash cloud surges, May 8 and 20, 1902. *Geology*, **8**, 472–6.

FISKE, R. S. 1963. Subaqueous pyroclastic flows in the Ohanapecosh Formation, Washington. *Bull. geol. Soc. Am.* **74**, 391–406.

—— & MATSUDA, T. 1964. Submarine equivalents of ash flows in the Tokiwa Formation, Japan. *Am. J. Sci.* **262**, 76–106.

——, HOPSON, C. A. & WATERS, A. C. 1963. Geology of Mount Rainier National Park, Washington. *Prof. Pap. U.S. geol. Surv.* **444**, 1–93.

FLEET, A. J. & MCKELVEY, B. C. 1978. Eocene explosive submarine volcanism, Ninetyeast Ridge, Indian Ocean. *Mar. Geol.* **26**, 73–97.

FRANCIS, E. H. & HOWELLS, M. F. 1973. Transgressive welded ash-flow tuffs among the Ordovician sediments of NE Snowdonia, North Wales. *J. geol. Soc. London*, **129**, 621–41.

FURNES, H., FRIDLEIFSSON, I. B. & ATKINS, F. B. 1980. Subglacial volcanics—on the formation of acid hyaloclastites. *J. Volcan. geoth. Res. Amsterdam*, **8**, 95–110.

GARRISON, R. E., ESPIRITU, E., HORAN, L. J. & MACK, L. E. 1979. Petrology, sedimentology, and diagenesis of hemipelagic limestone and tuffaceous turbidites in the Aksitero Formation, Central Luzon, Philippines. *Prof. Pap. U.S. geol. Surv.* 1112.

GLOPPEN, T. G. & STEEL, R. J. 1981. The deposits, internal structure and geometry in six alluvial fan–fan delta bodies (Devonian-Norway)—A study in the significance of bedding sequence in conglomerates. *Spec. Publ. Soc. econ. Paleontol. Mineral., Tulsa*, **31**, 49–69.

GONZALEZ-BONORINO, G. & MIDDLETON, G. V. 1976. A Devonian submarine fan in western Argentina. *J. sediment. Petrol.* **46**, 56–69.

HAHN, G. A., ROSE JR, W. I. & MEYERS, T. 1979. Geochemical correlation of genetically related rhyolitic ash-flow and air-fall ashes, central and western Guatemala and the equatorial Pacific. *Spec. Pap. geol. Soc. Am.* **180**, 101–12.

HAMPTON, M. A. 1972. The role of subaqueous debris flow in generating turbidity currents. *J. sediment. Petrol.* **42**, 775–93.

—— 1975. Competence of fine-grained debris flows. *J. sediment. Petrol.* **45**, 834–44.

——, BOUMA, A. H. & FROST, T. P. 1979. Volcanic ash in surficial sediments of the Kodiak shelf—an indicator of sediment dispersal patterns. *Mar. Geol.* **29**, 347–56.

HEIN, J. R., SCHOLL, D. W. & MILLER, J. 1978. Episodes of Aleutian Ridge explosive volcanism. *Science*, **199**, 137–41.

HISCOTT, R. N. & MIDDLETON, G. V. 1979. Depositional mechanics of thick bedded sandstones at the base of a submarine slope, Tourelle Formation (Lower Ordovician), Quebec, Canada. *Spec. Publ. Soc. econ. Paleontol. Mineral., Tulsa*, **27**, 307–26.

HONNOREZ, J. 1972. La Palagonitisation: l'alteration sous-marine du verre volcanique basique de Palagonia (Sicile). *Vukan-institut Immanuel Friedlaender* No. 9, *Birkhauser Verlag Basel und Stuttgart.*

HOWELLS, M. F. & LEVERIDGE, B. E. 1980. The Capel Curig Volcanic Formation. *Rep. Inst. Geol. Sci. Lond.* **80**(6), 1–23.

——, —— & EVANS, C. D. R. 1973. Ordovician ash-flow tuffs in eastern Snowdonia. *Rep. Inst. Geol. Sci. Lond.* **73**(3), 1–33.

——, ——, ADDISON, R., EVANS, C. D. R. & NUTT, M. J. C. 1979. The Capel Curig Volcanic Formation, Snowdonia, North Wales; variations in ash-flow tuffs related to emplacement environment. *In*: HARRIS, A. L., HOLLAND, C. H. & LEAKE, B. L (eds) *The Caledonides of the British Isles–Reviewed. Spec. Publ. geol. Soc. London*, **8**, 611–8. Blackwell Scientific Publications, Oxford.

HUANG, T. C., WATKINS, N. D. & SHAW, D. M. 1975. Atmospherically transported volcanic glass in deep-sea sediments: volcanism in sub-antarctic latitudes of the south Pacific during late Pleistocene time. *Bull. geol. Soc. Am.* **86**, 1305–15.

——, CAREY, S., SIGURDSSON, H. & DAVIS, A. 1979. Correlations and contrasts of deep-sea ash deposits from the Lesser Antilles in the Western Equatorial Atlantic and Eastern Caribbean at latitude 14°N. *Eos*, **59**, 1119.

WATKINS, N. D., SHAW, D. M. & KENNETT, J. P. 1973. Atmospherically transported volcanic dust in South Pacific deep sea sedimentary cores at distances over 3000 km from the eruptive source. *Earth planet. Sci. Lett.* **20**, 119–24.

INGERSOLL, R. V. 1978. Petrofacies and petrologic evolution of the Late Cretaceous forearc basin, northern and central California. *J. Geol.* **86**, 335–52.

JOHNSON, A. M. 1970. *Physical Processes in Geology.* Freeman, Cooper & Co., San Francisco.

JONES, J. G. 1970 Intraglacial volcanoes of the Laugarvatn region, southwest Iceland, II. *J. Geol.* **78**, 127–40.

KATO, I., MUROI, I., YAMAZAKI, T. & ABE, M. 1971. Subaqueous pyroclastic flow deposits in the upper Donzurubo Formation, Nijo-san district, Osaka, Japan. *J. Geol. Soc. Japan*, **77**, 193–206.

KELLER, J., RYAN, W. B. F., NINKOVICH, D. & ALTHERR, R. 1978. Explosive volcanic activity in the Mediterranean over the past 2000,000 yr as recorded in deep-sea sediments. *Bull. geol. Soc. Am.* **89**, 591–604.

KENNETT, J. P. 1981. Marine tephrochronolgy. *In*: EMILIANI, C. (ed.) *The Oceanic Lithosphere*, Vol. 7, John Wiley & Sons, Inc. Pp. 1373–436.

—— & THUNELL, R. C. 1975. Global increase in Quaternary volcanism. *Science*, **187**, 497–503.

—— & —— 1977. On explosive Cenozoic volcanism and climatic implications. *Science*, **196**, 1231–4.

KLEIN, G. DEV., OKADA, H. & MITSUI, K. 1979. Slope sediments in small basins associated with a Neogene active margin, western Hokkaido Island, Japan. *Spec. Publ. Soc. econ. Paleontol. Mineral., Tulsa*, **27**, 359–74.

KOKELAAR, B. P. 1982. Fluidization of wet sediments during the emplacement and cooling of various igneous bodies. *J. geol. Soc. London*, **139**, 21–33.

—— 1983. The mechanism of Surtseyan Volcanism. *J. geol. Soc. London*, **140**, 939–44.

KOMAR, P. D. 1971. Hydraulic jumps in turbidity currents. *Bull. geol. Soc. Am.* **82**, 1477–88.

KUENEN, PH. H. 1952. Estimated size of the Grand Banks turbidity current. *Am. J. Sci.* **250**, 874–84.

KUENZI, W. D., HORST, O. H. & MCGEHEE, R. V. 1979. Effect of volcanic activity on fluvial-deltaic sedimentation on a modern arc-trench gap, southwestern Guatemala. *Bull. geol. Soc. Am.* **90**, 827–38.

LACROIX, A. 1904. *La Montagne Pelée et Ses Eruptions*. Masson, Paris.

LAJOIE, J. 1979. Facies models 15. Volcaniclastic rocks. *Geosci. Can.* **6**, 129–39.

LEDBETTER, M. T. & SPARKS, R. S. J. 1979. Duration of large-magnitude explosive eruptions deduced from graded bedding in deep-sea ash layers. *Geology*, **7**, 240–4.

LEITCH, E. C. & CAWOOD, P. A. 1980. Olistoliths and debris flow deposits at ancient consuming plate margins: an eastern Australian example. *J. sediment. Petrol.* **25**, 5–22.

LEWIS, D. W. 1976. Subaqueous debris flows of early Pleistocene age at Motunau, North Canterbury, New Zealand. *N.Z. Jl Geol. Geophys.* **19**, 535–67.

LONSDALE, P. F. 1975. Sedimentation and tectonic modification of the Samoan archipelagic apron. *Bull. Am. Ass. Petrol. Geol.* **59**, 780–98.

—— 1977. Abyssal pahoehoe with lava coils at the Galapagos Rift. *Geology*, **5**, 147–52.

—— & SPIESS, F. N. 1979. A pair of young cratered volcanoes on the east Pacific rise. *J. Geol.* **87**, 157–73.

LOWE, D. R. 1976. Grain flow and grain flow deposits. *J. sediment. Petrol.* **46**, 188–99.

—— 1979. Sediment gravity flows: their classification and some problems of application to natural flows and deposits. *Spec. Publ. Soc. econ. Paleontol. Mineral., Tulsa*, **27**, 75–82.

—— 1982. Sediment gravity flows. II. Depositional models with special reference to the deposits of high-density turbidity currents. *J. sediment. Petrol.* **52**, 279–97.

LOWMAN, R. D. W. & BLOXAM, T. W. 1981. The petrology of the lower Paleozoic Fishguard Volcanic Group and associated rocks E of Fishguard, N Pembrokeshire (Dyfed), South Wales. *J. geol. Soc. London*, **138**, 47–68.

MACDONALD, G. A. 1972. *Volcanoes*. Prentice-Hall, New Jersey.

MANSFIELD, C. F. 1979. Upper Mesozoic subsea fan deposits in the southern Diablo Range, California: Record of the Sierra Nevada magmatic arc. *Bull. geol. Soc. Am.* **90**, 1025–46.

MCBIRNEY, A. R. 1963. Factors governing the nature of submarine volcanism. *Bull. Volcanol.* **26**, 455–69.

MCBRIDE, E. F. 1978. Olistostrome in the Tesnus Formation (Mississippian–Pennsylvanian), Payne Hills, Marathon region, Texas. *Bull. Geol. Soc. Am.* **89**, 1550–8.

MIDDLETON, G. V. 1967. Experiments on density and turbidity currents. III. Deposition of sediment. *Can. J. Earth Sci.* **4**, 475–505.

—— 1970. Experimental studies related to problems of flysch sedimentation. *Spec. Pap. geol. Ass. Can.* **7**, 253–72.

—— & HAMPTON, M. A. 1973. Sediment gravity flows: Mechanics of flow and deposition. *In*: MIDDLETON, G. V. & BOUMA, A. H. (eds) *Turbidites and Deep-water Sedimentation.* Short Course, Soc. Econ. Paleont. Mineral. Pacific Section, May 1973, pp. 1–38.

—— & —— 1976. Subaqueous sediment transport and deposition by sediment gravity flows. *In*: STANLEY, D. J. & SWIFT, D. J. P. (eds) *Marine Sediment Transport and Environmental Management.* John Wiley & Sons, pp. 197–218.

MITCHELL, A. H. G. 1970. Facies of an early Miocene volcanic arc, Malekula Island, New Hebrides. *Sedimentology*, **14**, 201–43.

—— & READING, H. G. 1978. Sedimentation and tectonics. *In*: READING, H. G. (ed.) *Sedimentary Environments and Facies*, **14**, 439–76. Blackwell Scientific Publications, Oxford.

MOORE, J. G. 1965. Petrology of deep-sea basalt near Hawaii. *Am. J. Sci.* **263**, 40–52.

—— 1970. Water content of basalt erupted on the ocean floor. *Contrib. Mineral. Petrol.* **28**, 272–9.

—— 1975. Mechanism of formation of pillows. *Am. Sci.* **63**, 269–77.

—— & FISKE, R. S. 1969. Volcanic substructure inferred from dredge samples and ocean-bottom photographs, Hawaii. *Bull. geol. Soc. Am.* **80**, 1191–202.

—— & SCHILLING, J.-G. 1973. Vesicles, water, and sulfur in Reykjanes Ridge basalts. *Contrib. Mineral. Petrol.* **41**, 105–18.

—— & SISSON, T. W. 1981. Deposits and effects of the May 18 pyroclastic surge. *In*: LIPMAN, P. W. & MULLINEAUX, D. R. (eds) *The 1980 Eruptions of Mount St Helens, Washington. Prof. Pap. U.S. geol. Surv.* **1250**, 421–38.

MORGENSTERN, N. R. 1967. Submarine slumping and the initiation of turbidity currents. *In*: RICHARDS, A. F. (ed.) *Marine Geotechnique*, pp. 189–220.

NAKAMURA, K. 1974. Preliminary estimate of global volcanic production rate. *In*: FURUMOTO, S., MINIKAMI, T. & YUHARA, K. (eds) *The Utilization of Volcano Energy.* The Proceedings of U.S.–Japan Cooperative Science Seminar, Sandia Laboratories, Albuquerque, New Mexico. Pp. 273–84.

NARDIN, T. R., HEIN, F. J., GORSLINE, D. S. & EDWARDS, B. D. 1979. A review of mass movement processes, sediment and acoustic characteristics, and contrasts in slope and base-of-slope systems *versus* canyon–fan–basin floor systems. *Spec. Publ. Soc. econ. Paleontol. Mineral., Tulsa*, **27**, 61–73.

NAYLOR, M. A. 1980. The origin of inverse grading in muddy debris flow deposits—a review. *J. sediment. Petrol.* **50**, 1111–6.

—— 1981. Debris flow (olistostromes) and slumping on a distal passive continental margin: the Palombini limestone-shale sequence of the northern Apennines. *Sedimentology*, **28**, 837–52.

NAYUDU, Y. R. 1964. Volcanic ash deposits in the Gulf of Alaska and problems of correlation of deep-sea ash deposits. *Mar. Geol.* **1**, 194–212.

NEEDHAM, J. D. & FRANCHETEAU, J. 1974. Some characteristics of the rift valley in the Atlantic Ocean near 36°48′N. *Earth planet. Sci. Lett.* **22**, 29–43.

NEMEC, W., POREBSKI, S. J. & STEEL, R. J. 1980.

Texture and structure of resedimented conglomerates: examples from Ksiaz Formation (Famenian–Tournasian), southwestern Poland. *Sedimentology*, **27**, 519–38.

Niem, A. R. 1977. Mississippian pyroclastic flow and ash-fall deposits in the deep-marine Ouachita flysch basin, Oklahoma and Arkansas. *Bull. geol. Soc. Am.* **88**, 49–61.

Ninkovich, D. 1979. Distribution, age and chemical composition of tephra layers in deep-sea sediments off western Indonesia. *J. Volcan. geoth. Res. Amsterdam*, **5**, 67–86.

—— & Donn, W. L. 1976. Explosive Cenozoic volcanism and climatic implications. *Science*, **194**, 899–906.

—— & Heezen, B. C. 1965. Santorini tephra. *In*: *Submarine Geology and Geophysics* (Colston Papers No. 17), **27**, 413–53. Butterworths Scientific Publications, London.

—— & —— 1967. Physical and chemical properties of volcanic glass shards from Pozzolana ash, Thera Island, and from upper and lower ash layers in eastern Mediterranean deep-sea cores. *Nature, Lond.* **213**, 582–4.

—— & Shackleton, N. J. 1975. Distribution, stratigraphic position and age of ash layer 'L', in the Panama Basin region. *Earth planet. Sci. Lett.* **27**, 20–34.

——, Sparks, R. S. J. & Ledbetter, M. T. 1978. The exceptional magnitude and intensity of the Toba eruption, Sumatra: An example of the use of deep-sea tephra layers as a geological tool. *Bull. Volcanol.* **41-3**, 286–98.

Peterson, D. W. 1976. Processes of volcanic island growth, Kilauea Volcano, Hawaii, 1969–1973. *In*: Ferran, O. G. (ed.) *Proceedings of the Symposium 'Andean and Antarctic Volcanology Problems'*. Spec. Series Internat. Assoc. of Volcanology and Chemistry of the Earth's Interior, 172–89.

Riehle, J. R. 1973. Calculated compaction profiles of rhyolitic ash-flow tuffs. *Bull. geol. Soc. Am.* **84**, 2193–216.

Rittmann, A. 1962. *Volcanoes and Their Activity.* John Wiley and Sons, New York.

Robinson, P. T., Flower, M. F. J., Swanson, D. A. & Staudigel, H. 1980. Lithology and eruptive stratigraphy of Cretaceous oceanic crust, western Atlantic ocean. *Init. Rep. Deep Sea drill. Proj.* **51-3**, 1535–55.

Scheidegger, K. F., Jezek, P. A. & Ninkovich, D. 1978. Chemical and optical studies of glass shards in Pleistocene and Pliocene ash layers from DSDP Site 192, northwest Pacific Ocean. *J. Volcan. geoth. Res. Amsterdam*, **4**, 99–116.

——, Corliss, J. B., Jezek, P. A. & Ninkovich, D. 1980. Composition of deep-sea ash layers derived from north Pacific volcanic arcs: variations in time and space. *J. Volcan. geoth. Res. Amsterdam*, **7**, 107–37.

Schmincke, H.-U. 1967. Graded lahars in the type section of the Ellensburg Formation, south-central Washington. *J. sediment. Petrol.* **37**, 438–48.

—— & von Rad, U. 1979. Neogene evolution of Canary Island volcanism inferred from ash layers and volcaniclastic standstones of DSDP Site 397 (Leg 47A). *Init. Rep. Deep Sea drill. Proj.* **47**(1), 703–25.

Self, S. & Rampino, M. R. 1981. The 1883 eruption of Krakatau. *Nature, Lond.* **294**, 699–704.

Shaw, D. M., Watkins, N. D. & Huang, T. C. 1974. Atmospherically transported volcanic dust in deep sea sediments: theoretical consideration. *J. geophys. Res.* **79**, 3087–97.

Sheridan, M. F. 1979. Emplacement of pyroclastic flows: a review. *Spec. Pap. geol. Soc. Am.* **180**, 125–36.

Sigurdsson, H., Sparks, R. S. J., Carey, S. N. & Huang, T. C. 1980. Volcanogenic sedimentation in the Lesser Antilles Arc. *J. Geol.* **88**, 523–40.

Smith, R. L. 1960. Ash flows. *Bull. geol. Soc. Am.* **71**, 795–842.

Sparks, R. S. J. 1976. Grain size variations in ignimbrites and implications for the transport of pyroclastic flows. *Sedimentology*, **23**, 147–88.

——, Sigurdsson, H. & Carey, S. 1980a. The entrance of pyroclastic flows into the sea. I. Oceanographic and geologic evidence evidence from Dominica. Lesser Antilles. *J. Volcan. geoth. Res. Amsterdam*, **7**, 87–96.

——, —— & —— 1980b. The entrance of pyroclastic flows into the sea. II. Theoretical considerations on subaqueous emplacement and welding. *J. Volcan. geoth. Res. Amsterdam*, **7**, 97–105.

Stanley, D. J. & Taylor, P. T. 1977. Sediment transport down a seamount flank by a combined current and gravity process. *Mar. Geol.* **23**, 77–88.

——, Palmer, H. D. & Dill, R. F. 1978. Coarse sediment transport by mass flow and turbidity current processes and downslope transformations in Annot Sandstone Canyon–Fan Valley systems. *In*: Stanley, D. J. & Kelling, G. (eds) *Sedimentation in Submarine Canyons, Fans and Trenches*. Dowden, Hutchinson & Ross, Stroudsburg, Penn. 85–115.

Stanton, W. I. 1960. The lower Paleozoic rocks of south-west Murrisk, Ireland. *J. geol. Soc. London*, **116**, 269–96.

Staudigel, H. & Schmincke, H.-U. 1981. Structural evolution of a seamount: evidence from the uplifted intraplate seamount on the island of La Palma, Canary Islands. *Eos*, **62**, 1075.

Swarbrick, R. E. & Naylor, M. A. 1980. The Kathikas melange, SW Cyprus, Cretaceous submarine debris flows. *Sedimentology*, **26**, 63–78.

Tasse, N., Lajoie, J. & Dimroth, E. 1978. The anatomy and interpretation of an Archean volcaniclastic sequence, Noranda region, Quebec. *Can. J. Earth Sci.* **15**, 874–88.

Thunell, R., Federman, A., Sparks, S. & Willams, D. 1979. The age, origin and volcanological significance of the Y-5 ash layer in the Mediterranean. *Quat. Res.* **12**, 241–53.

Van Andel, Tj. H. & Komar, P. D. 1969. Ponded sediments of the Mid-Atlantic Ridge between 22° and 23° North latitude. *Bull. geol. Soc. Am.* **80**, 1163–90.

VESSEL, R. K. & DAVIES, D. K. 1981. Nonmarine sedimentation in an active forre arc basin. *Spec. Publ. Soc. econ. Paleontol. Mineral., Tulsa*, **31**, 31–45.

WALKER, G. P. L. 1979. A volcanic ash generated by explosions where ignimbrite entered the sea. *Nature, Lond.* **281**, 642–6.

WATKINS, N. D. & HUANG, T. C. 1977. Tephras in abyssal sediments east of the North Island, New Zealand: Chronology, paleowind velocity, and paleoexplosivity. *N.Z. Jl Geol. Geophys.* **20**, 179–98.

——, SPARKS, R. S. J., SIGURDSSON, H., HUANG, T. C., FEDERMAN, A., CAREY, S. & NINKOVICH, D. 1978. Volume and extent of the Minoan tephra from Santorini Volcano: new evidence from deep-sea sediment cores. *Nature, Lond.* **271**, 122–6.

WILSON, C. J. N. 1980. The role of fluidization in the emplacement of pyroclastic flows: An experimental approach. *J. Volcan. geoth. Res. Amsterdam*, **8**, 231–49.

WOHLETZ, K. H. 1980. Explosive Hydromagmatic Volcanism. Thesis, Ph.D. Arizona State University (unpub.).

—— & SHERIDAN, M. F. 1979. A model of pyroclastic surge. *Spec. Pap. geol. Soc. Am.* **180**, 177–94.

WRIGHT, J. V. & MUTTI, E. 1981. The Dali Ash, Island of Rhodes, Greece: a problem in interpreting submarine volcanigenic sediments. *Bull. Volcanol.* **44–2**, 153–67.

YAMADA, E. 1973. Subaqueous pumice flow deposits in the Onikobe Caldera, Miyagi Prefecture, Japan. *J. geol. Soc. Japan*, **79**, 589–97.

YAMAZAKI, T., KATO, I., MUROI, I. & ABE, M. 1973. Textural analysis and flow mechanism of the Donzurubo subaqueous pyroclastic flow deposits. *Bull. Volcanol.* **37**, 231–44.

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