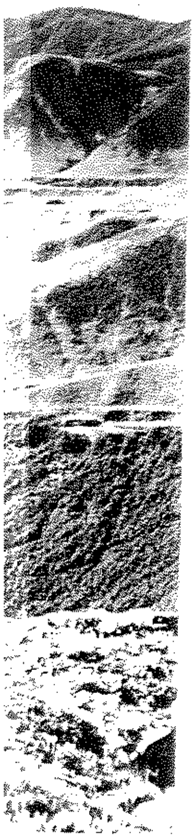

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CHAPTER SEVEN

# *Transport and deposition of subaerial pyroclastic flows and surges*

## Initial statement

In this chapter we examine more closely the transportational and depositional processes associated with pyroclastic flows and surges (Ch. 5). Both are a type of flow, one at the dilute, low particle concentration end of the spectrum and the other at the highly concentrated end. It is shown that there is a broad spectrum of pyroclastic flow types, and that the flow mechanics can vary, as do the resultant facies. The roles of fluidisation and turbulence are evaluated. The anatomy of pyroclastic flows is also examined, and consideration is given to the characteristic depositional processes associated with the head, body and trailing cloud. We similarly look in detail at the flow and depositional mechanics associated with surges, and consider the differences between wet and dry surges. Finally, we look at the possible relationships and differences between pyroclastic flows and surges.

## 7.1 Subaerial pyroclastic flows as high particle concentration flows

All workers would now agree that *pyroclastic flows are gravity controlled* and tend to move along topographic depressions. However, there has been, and still is, much debate on the details of flow mechanisms. Because of poor sorting, earlier workers thought turbulence within flows was important (R. L. Smith 1960a, Murai 1961, Fisher 1966a). The ability of some flows to surmount topographic barriers (Plate 7) led some workers to suggest that they were greatly expanded, implying dilute, turbulent flows (Yokoyama 1974, Sheridan & Ragan 1976). In contrast, from a detailed study of grainsize variations in ignimbrites, Sparks (1976) postulated that *pyroclastic flows are high concentration, poorly expanded, partially fluidised flows*, and are in many ways analogous to types of debris flow in which poor sorting is attributed to high particle concentration, not turbulence (Chs 2 & 10). Their transport and depositional mechanisms may be similar to debris flows, but in debris flows larger clasts are carried by a matrix of mud and water (cohesive flows) or of poorly sorted granular material and water (grain dominant flows), whereas in pyroclastic flows they are carried by fine ash and gas. Sparks (1976) concluded that many pyroclastic flows may be laminar in their movement, especially in the body. Several more-recent studies have concurred with this view (Sparks *et al.* 1978, Sheridan 1979, J. V. Wright & Walker 1977, 1981, C. J. N. Wilson 1980), although it is accepted that variable degrees of turbulence may be important in the head regions, and even in the bodies of violent flows.

Pyroclastic flows encompass a wide range of phenomena. At the one extreme are the small denser-clast flows frequently observed in historic eruptions (Ch. 5). These may come to rest on substantial slopes, transport very large blocks and have well developed levées (Fig. 7.1a) and flow fronts. All of these features suggest that these flows had a substantial yield strength, and that deposition was almost instantaneous by *en masse* freezing of the flows, as is generally believed to be the case in epiclastic debris flows (A. M. Johnson 1970; Ch. 10). Such flows act as simple avalanches of debris, and gas probably plays only a minor role in their movement. They are likely to flow by a combination of laminar and plug flow, typical of Bingham fluids with a high yield strength (Ch. 2). At the other extreme are the pumice flows which form ignimbrites. These generally do not transport excessively large blocks for great distances, they are generally not found on steep slopes and neither levées, nor any other surface depositional features are not commonly observed. However, pumice flows still produce poorly sorted, structureless deposits, and all of the evidence indicates they move as high

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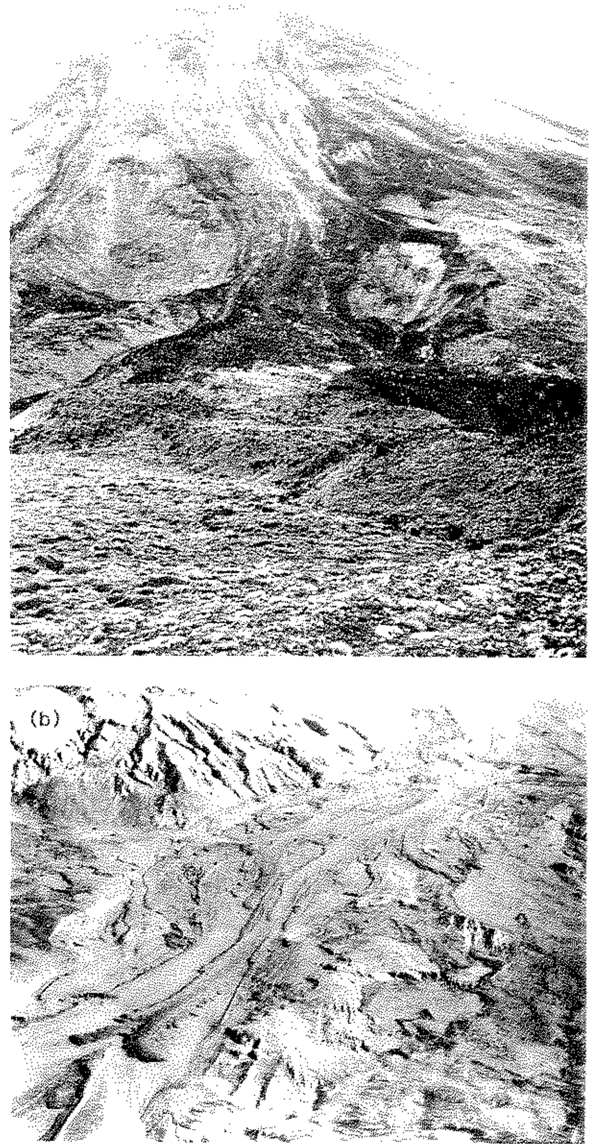

(b)

**Figure 7.1** (a) Scoria-flow deposits from the 1975 eruption of Ngauruhoe with well developed levées and surface ridges. (b) Oblique view of pumice flow lobes from the 22 July and 7 August 1980, eruptions of Mt St Helens, showing their well defined surface morphology and the generally constant width and thickness of the deposits. (After L. Wilson & Head 1981.)

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particle concentration flows. However, it is now recognised that there is a complete spectrum of ignimbrite types. Some of the Mt St Helens pumice-flow deposits (22 July and 7 August flows), although these produced only very small volume deposits ($<0.001$ km$^3$), have levées and surface features (Fig. 7.1b), indicating that the flows may have had higher yield strengths than are generally expected for pumice flows. Extremely *violent* pumice flows (C. J. N. Wilson & Walker 1981, G. P. L. Walker & Wilson 1983), can leave a topography mantling deposit and transport larger lithic blocks considerable distances. The best-described example is from the Taupo ignimbrite (Section 7.5; Ch. 8), which is very different in many respects from currently accepted, '*conventional*' ignimbrites.

Large pumice flows can travel for distances of tens of kilometres, and their mobility has been considered spectacular because of their ability to surmount topographic barriers. For example, the Ito pyroclastic flow (Yokoyama 1974) must have surmounted a 600 m high mountain pass 60 km from source (Fig. 8.3) and the Fisher ash flow tuff (Miller & Smith 1977) crossed a 500 m barrier 25 km from source. As mentioned previously,

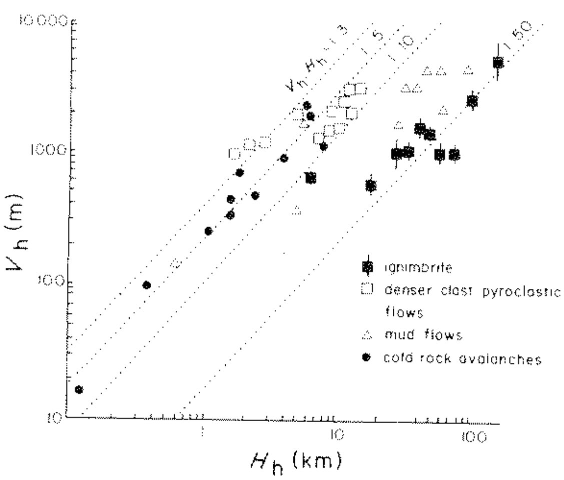


**Figure 7.2** Relationship between the vertical height dropped ($V_h$) and the horizontal distance travelled ($H_h$) for types of pyroclastic flow and debris flow. $V_h$ for ignimbrites is usually taken as the reconstructed height of the volcano before caldera collapse, and bars indicate the uncertainty. The data indicate that ignimbrites are apparently more mobile than other types of debris flow. (After Sparks 1976.)

some workers accounted for this mobility by suggesting that pumice flows are highly expanded. However, given sufficient momentum, a high concentration pumice flow could surmount such barriers, and indeed cold rock avalanches (Ch. 10) are known to travel uphill, e.g. the Saidmarreh landslide in Iran climbed over a 600 m barrier (P. E. Kent 1966). If one allows for eruption column collapse from heights of several kilometres or more, pumice flows are found to be no less mobile than other types of mass flow (Fig. 7.2). Measurements of the heights climbed by a pyroclastic flow can be used to estimate minimum palaeoflow velocities from the simple potential energy to kinetic energy relationship

$$gh = v^2 \qquad (7.1)$$

where $h$ is the height climbed, $v$ is the velocity and $g$ is the acceleration due to gravity. Absolute-minimum velocities of 60–160 m s$^{-1}$ are inferred from several ignimbrites (e.g. Yokoyama 1974, Sparks 1976, P. W. Francis & Baker 1977, Miller & Smith 1977, Barberi *et al.* 1978). The Saidmarreh landslide must have had a velocity of at least 100 m s$^{-1}$, and thus such high flow velocities are not unique to pyroclastic flows.

## 7.2 Fluidisation

Although the mobility and momentum of pyroclastic flows can be attributed in the first instance to either the momentum acquired during collapse from a high eruption column (cf. rock avalanches, Ch. 10), or the high eruption rates and associated high exit velocities; this mobility can be enhanced by the inclusion of a lubricating fluid within the flow, especially if that fluid also provides dynamic support or uplift to the grain population, or part of it, during flow. In that way, the fluid would retard sedimentation from the flow, and so act to reduce the frictional interaction between the flow and the substrate.

Fluidisation is commonly believed to play an important role in this regard, in the transport of pyroclastic flows. As an industrial process, fluidisation was developed largely during and immediately after World War II. D. L. Reynolds (1954) was the first geologist to examine the chemical engineering literature and suggest that fluidisation might have geological applications, including pyroclastic flows. More recently, its role in the emplacement of pyroclastic flows has been discussed by McTaggart (1960), Sparks (1976, 1978b), Sheridan (1979) and C. J. N. Wilson (1980, 1984).

When an upward stream of gas (or liquid) is passed at increasing velocity ($U$) through a bed of cohesionless particulate solids, fluidisation is the condition attained at a certain critical fluid velocity (the *minimum fluidisation velocity*, $U_{mf}$) when the

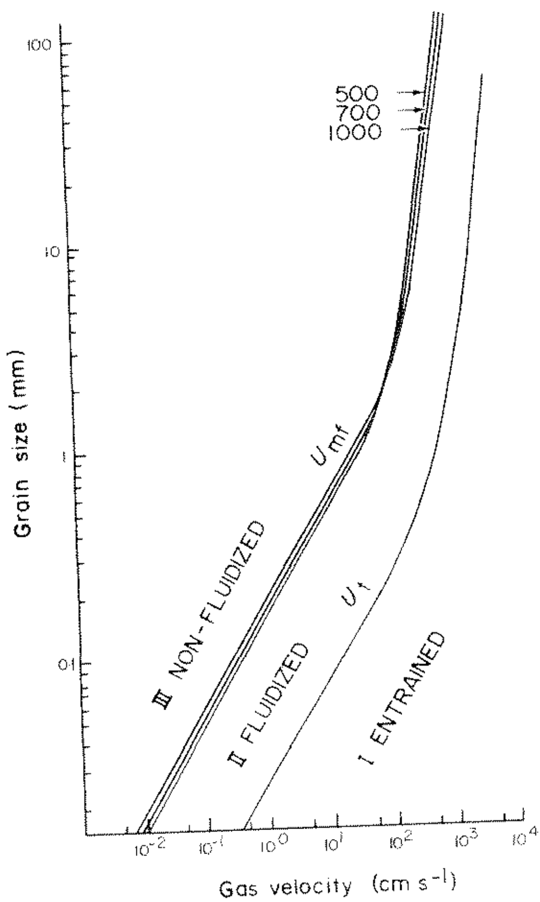


**Figure 7.3** Theoretical curves of the relationship of minimum fluidisation velocity ($U_{mf}$) and terminal fall velocity ($U_t$) and grainsize. Curves are calculated for spheres of density 1.0 g cm$^{-3}$ in $CO_2$ with a voidage of 0.45 at 500, 700 and 1000°C. The grainsize–$U_t$ curve is calculated at 700°C for $CO_2$. (After Sparks 1976.)

drag force exerted across the bed by the fluid is equal to the buoyant weight of the bed (see C. J. N. Wilson 1980, 1984). In this state the bed no longer exists as a coherent mass, but takes on a fluid-like character. However, it is more appropriate here (see below) to use the term in a looser sense (used by chemical engineers) to cover all conditions, from very low gas velocities and packed-bed conditions ($0 < U < U_{mf}$) to high flow velocities and dilute-phase fluidisation ($U >> U_{mf}$).

Sparks (1976) showed theoretically that pyroclastic flows can only be semi-fluidised. In polydispersive systems (grain populations with a wide range of grainsizes or densities, or both), such as in pyroclastic flows, before the largest particles become fluidised the gas flow velocity exceeds the terminal fall velocity ($U_t$) of the smallest sizes. Such particles are entrained by the gas and carried out and lost from the system. This process is called **elutriation**, and is important industrially and in the transport of pyroclastic flows. Sparks calculated curves of $U_{mf}$ and $U_t$ for a wide range of particles and conditions likely to occur in pyroclastic flows (Fig. 7.3). These calculations used formulae from standard chemical engineering literature (e.g. Kunii & Levenspiel 1969). $U_{mf}$ can be determined from the so-called modified Ergun equation:

$$\frac{1.75}{Q_s e_{mf}^3}\left(\frac{d_p U_{mf} \varrho_g}{\eta}\right)^2 + \frac{150(1-e_{mf})}{Q_s^2 e_{mf}^3}\frac{d_p U_{mf} \varrho_g}{\eta} = \frac{d_p^3 \varrho_g(\varrho_s - \varrho_g)g}{\eta^2} \quad (7.2)$$

where $U_{mf}$ is the minimum velocity of fluidisation (cm s$^{-1}$), $\varrho_g$ and $\varrho_s$ are the densities of gas and solid, respectively (g cm$^{-3}$), $d_p$ is mean particle diameter (cm), $Q_s$ is the sphericity of the particle (dimensionless) (= surface area of sphere/surface area of particle where sphere and particle have same volume), $e_{mf}$ is the voidage per unit volume at minimum fluidisation state, $\eta$ is the viscosity of gas (poise = $10^{-1}$ Pa s) and $g$ is the acceleration due to gravity.

$U_t$ for a particle in a fluid is given by

$$U_t = \frac{g(\varrho_s - \varrho_g)d_p^2}{18\eta}, \qquad \text{when Re} < 0.4 \quad (7.3$$

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$$U_t = \left(\frac{4(\rho_s - \rho_g)^2 g^2}{225 \rho g \eta}\right)^{1/3} d_p,$$

when $0.4 < \text{Re} < 500$ (7.4)

$$U_t = \left(3.1g\,\frac{(\rho_s - \rho g)^2}{\rho}\,d_p\right)^{1/2}, \text{ when Re} > 500 \qquad (7.5)$$

where Re is the Reynolds Number (Eqn 2.6: Ch. 2).

Figure 7.3 shows that for any gas velocity it is only possible to fluidise a limited range of grainsizes that satisfy the condition $U_t > U > U_{mf}$. This led Sparks to suggest that in pyroclastic flows there must always be three phases when gas is passing through:

*Phase I*: particles with $U_t < U$,
*Phase II*: particles with $U_t \geqslant U \geqslant U_{mf}$ and
*Phase III*: particles with $U_{mf} > U$.

A pyroclastic flow was envisaged to comprise a matrix of particles of phases I and II, in which is dispersed particles of phase III which tend to float (pumice) or sink (lithics) depending on the density contrast with the matrix (see Section 7.3.3). Phase I particles are lost by elutriation from the matrix to form the dilute overriding ash cloud (see Fig. 5.13).

Later experimental work by Sparks (1978b), Sheridan (1979) and C. J. N. Wilson (1980, 1984) also demonstrated that pumice flows and pyroclastic

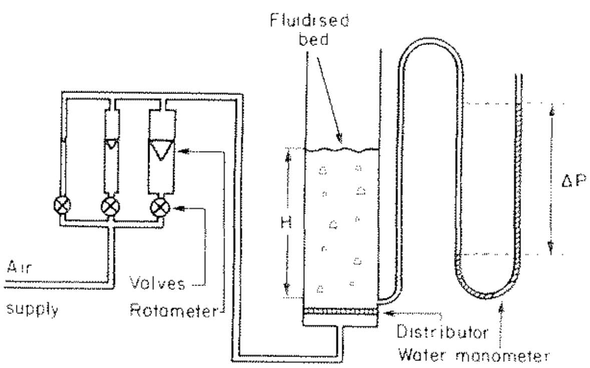



**Figure 7.4** Schematic diagram of a fluidisation rig. A high pressure air supply is regulated with valves, through rotameters and the distributor, into the fluidised bed. During runs, the height of the bed (*H*) and the gas flow rates are recorded, together with the pressure drop ($\Delta P$) on a water manometer. (After C. J. N. Wilson 1980.)

flows in general can only be semi-fluidised. The most comprehensive studies are those by C. J. N. Wilson (1980, 1984). Wilson reports fluidisation experiments on simple systems and on ignimbrite samples, and has supported his arguments with an extensive search into the chemical engineering literature. His results show that the fluidisation behaviour of pyroclastic flow samples differ radically from any simple system, principally because of the variable hydraulic properties of the different grain types and their resultant poor sorting.

A fluidisation rig of the type used by Wilson is illustrated in Figure 7.4. Examples of fluidisation plots of $U$ versus $\Delta P/H$ (the pressure drop across the bed per unit thickness of bed) are given in Figure 7.5. For narrow grainsize populations of ideally smooth and spherical particles, a fluidisation plot shows two straight lines which intersect at, and define, $U_{mf}$ (Fig. 7.5a). (The slope of the straight line for $U < U_{mf}$ can be predicted by reference to published correlations, and this information has been used in the modified Ergun equation to obtain correlations of $U_{mf}$ versus particle and fluid characteristics.) Commonly, some degree of hysteresis is evident between the curve corresponding to increasing $U$ and the one corresponding to decreasing $U$ (Fig. 7.5b). This can be related to voidage changes during fluidisation, a wide grainsize distribution and/or irregular particle shape. For materials having a wide grainsize variation and irregular particle shapes, a plot is obtained with gross hysteresis (Fig. 7.5c).

Poorly sorted ($\sigma_{\phi} > 1.0$) materials, such as sand mixes and pyroclastic flow samples, show distinctive fluidisation behaviour (Figs 7.5c & 6). At a certain gas velocity, designated $U_{ie}$ (the value of which cannot be predicted), some samples begin to expand, whereas at higher gas velocities, designated $U_{mp}$, samples begin to show segregation structures (Fig. 7.7). $U_{mp}$ is the gas velocity at the maximum pressure-drop that can be sustained across the bed (i.e. where $\Delta P/H$ is equal to the buoyant bulk density of the bed multiplied by the acceleration due to gravity: Fig. 7.6a). $U_{mp}$ replaces $U_{mf}$ measured in simple systems, but cannot be predicted reliably from published $U_{mf}$ correlations (C. J. N. Wilson 1984).

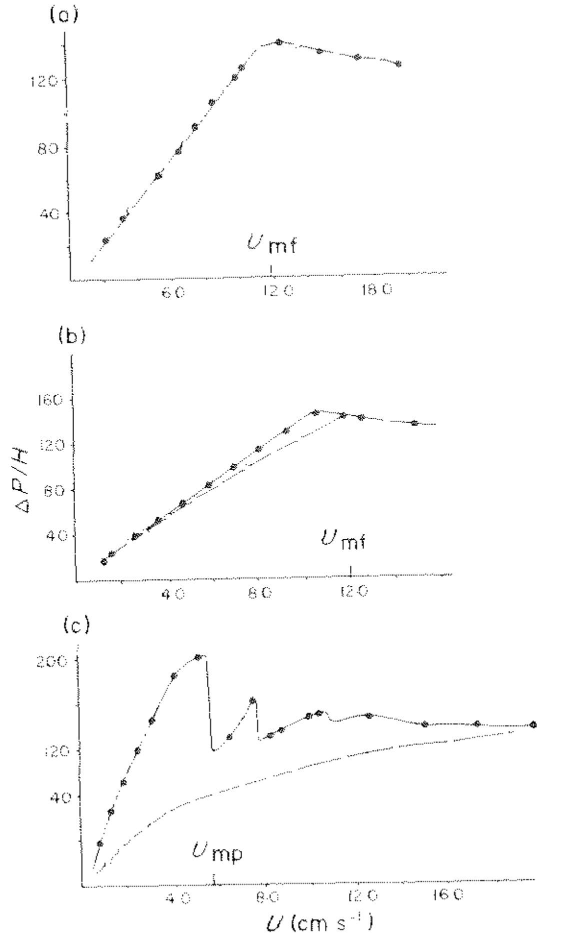



**Figure 7.5** Fluidisation plots for several materials. $\Delta P/H$ is in centimetres of water per metre of bed thickness; filled circles define the runs with increasing $U$ and open circles the runs with decreasing $U$. See text for the definitions of $U$, $U_{mf}$, $U_{mp}$ and $\Delta P/H$. (a) Glass spheres, $Md_{\phi}$ = 1.40, $\sigma_{\phi}$ = 0.22. (b) Quartz sand, $Md_{\phi}$ = 1.68, $\sigma_{\phi}$ = 0.32. (c) $Md_{\phi}$ = 1.28, $\sigma_{\phi}$ = 1.55. (After C. J. N. Wilson 1980.)

The up-curve on fluidisation plots of pyroclastic flow samples can therefore be divided into three sections:

*Section 1*: $0 < U < U_{ie}$,
*Section 2*: $U_{ie} \leq U \leq U_{mf}$ and
*Section 3*: $U > U_{mf}$.

In section 1, no expansion of the bed occurs and there is no loss of fines by elutriation (except perhaps from the surface layer of the bed). In section 2, although superficially similar to section 1, the bed has partially fluidised and expanded to accommodate the gas flow. Elutriation of fines is still absent. Section 3 is where an instability sets in, and part of the gas flow is concentrated into discrete channels. This leads to a sharp reduction in the $\Delta P/H$ ratio, which is accompanied by bubbling, the deposition of coarser or coarser and denser material in the channels (Figs 7.7a & b) and the elutriation of fines. At higher gas velocities, bubble-induced circulation destroys the pipes formed earlier, and irregular segregation pods settle towards the base of the bed (Fig. 7.7c). Gradually, well

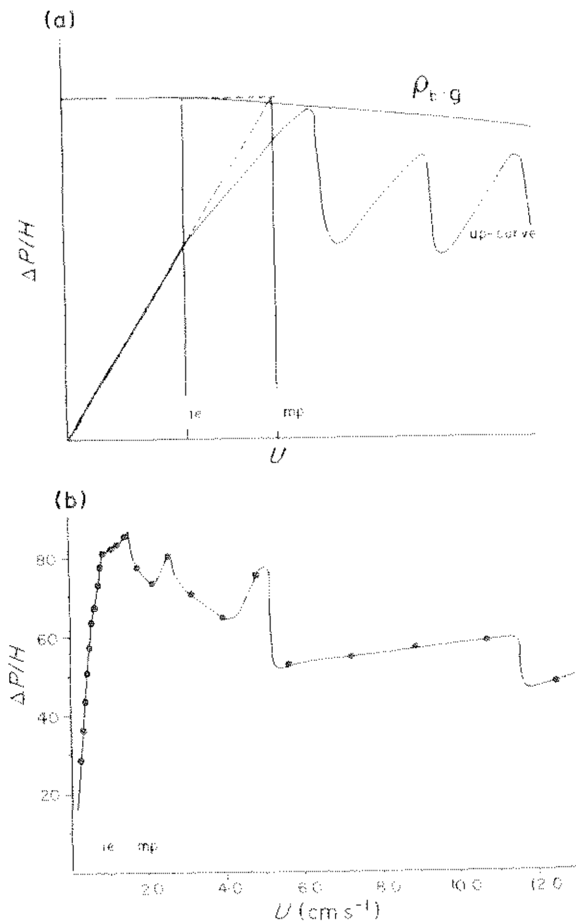



**Figure 7.6** (a) Typical curve with $U$ increasing for a poorly sorted sample such as of a pyroclastic flow to show how $U_{ie}$ and $U_{mf}$ are defined. (b) Curve with increasing $U$ for an actual sample of ignimbrite fines ($Md_{\phi}$ = 1.22, $\sigma_{\phi}$ = 1.75). (After C. J. N. Wilson 1980, 1984.)

defined segregation layers form at the top or bottom (or both) of the bed (Figs 7.7d & e). Finer and lighter particles move to the top of the bed, while coarser and denser particles sink to the base. In extreme cases, the bed may become completely gas-sorted.

C. J. N. Wilson (1980) proposed that the fluidisation behaviour of all pyroclastic flows can be typified by a fluidisation plot similar to those in Figure 7.6. This was used as the basis for a classification of pyroclastic flow types. Types 1, 2 and 3 (Table 7.1, Fig. 7.8) were introduced to relate a pyroclastic flow to the corresponding section on a fluidisation plot. The different kinds of grading and the causative mechanisms found in pyroclastic flows are discussed further below (Section 7.3). From Wilson's results, it is evident that only in type 3 flows can the processes described by Sparks (1976) operate freely.

Rheologically, fluidised systems behave in a non-Newtonian manner. Data in the chemical engineering literature deal almost exclusively with well sorted materials ($\sigma_\phi < 1.0$) at high gas velocities ($U > U_{mf}$). Under such conditions these materials

**Table 7.1** Classification of pyroclastic flow types based on fluidisation behaviour (after C. J. N. Wilson 1980).

| Flow type | Fluidisation behaviour with increasing gas flow velocities | Description |
|---|---|---|
| 1 | non-expanded | Non-graded flows; lack of expansion and high yield strength inhibits gravitational coarse-tail grading. If grading of larger clasts is present, it is due to some other mechanism. Minimal loss of ash into accompanying ash-cloud deposits |
| 2 | expanded | Expansion of flow allows gravitational coarse-tail grading of pumice and lithic clasts |
| 3 | segregating | High degree of fluidisation results in extreme grading. Distinct concentration zones of pumice and lithics are found, and other gas segregation structures. Large volume of vitric ash lost into co-ignimbrite fall |

have a non-linear relationship of stress to strain rate, and a negligible yield strength. C. J. N. Wilson (1980) considered the rheology of fluidised systems at $U < U_{mf}$ as follows. At rest ($U = 0$) the bed behaves as a particulate material which, at a given depth in the bed, can support a certain differential stress before failure occurs at a stage when the yield strength is exceeded. The yield strength ($S_0$) increases with depth:

$$S_0 = \mu \varrho g d \tag{7.6}$$

where $\mu$ is the tangent of the internal angle of friction, $\varrho$ is the buoyant bulk density of the material, $g$ is the acceleration due to gravity and $d$ is the depth in the bed. In a partly fluidised bed ($0 < U \leqslant U_{mf}$), the passage of gas results in a pressure drop across the bed, reducing the yield strength at a given depth. At $U = U_{mf}$ the yield strength is effectively zero and, from $0 < U \leqslant U_{mf}$, the yield strength in general ($S_u$) varies as:

$$S_u = (1 - U/U_{mf})S_0 \tag{7.7}$$

$$= (1 - U/U_{mf})\mu \varrho g d \tag{7.8}$$

Once full bed support is achieved, materials with $\sigma_\phi < 1$ may be expected to have similar rheologies to published industrial examples. For materials with $\sigma_\phi > 1$, the rheology at high gas flow rates ($U > U_{mp}$) is more complex. Much of the gas flow is diverted through segregation channels in which the material is better sorted, and this means that although the bed as a whole has a higher permeability, it continues to have a yield strength.

The rheology of fluidised materials in pyroclastic flows is therefore liable to be very complex. But from the foregoing discussion we can surmise that fluidisation will effectively reduce the yield strength of a flow. This concurs with field observations of the morphological and internal features of deposits. Poorly fluidised type 1 flows show features which indicate they had high yield strengths, and a Bingham model may approximate their motion. Yield strength is an important control of the types of grading observed in flows, and highly fluidised flows have very different characteristics. Fluidisation will also induce a stable density stratification

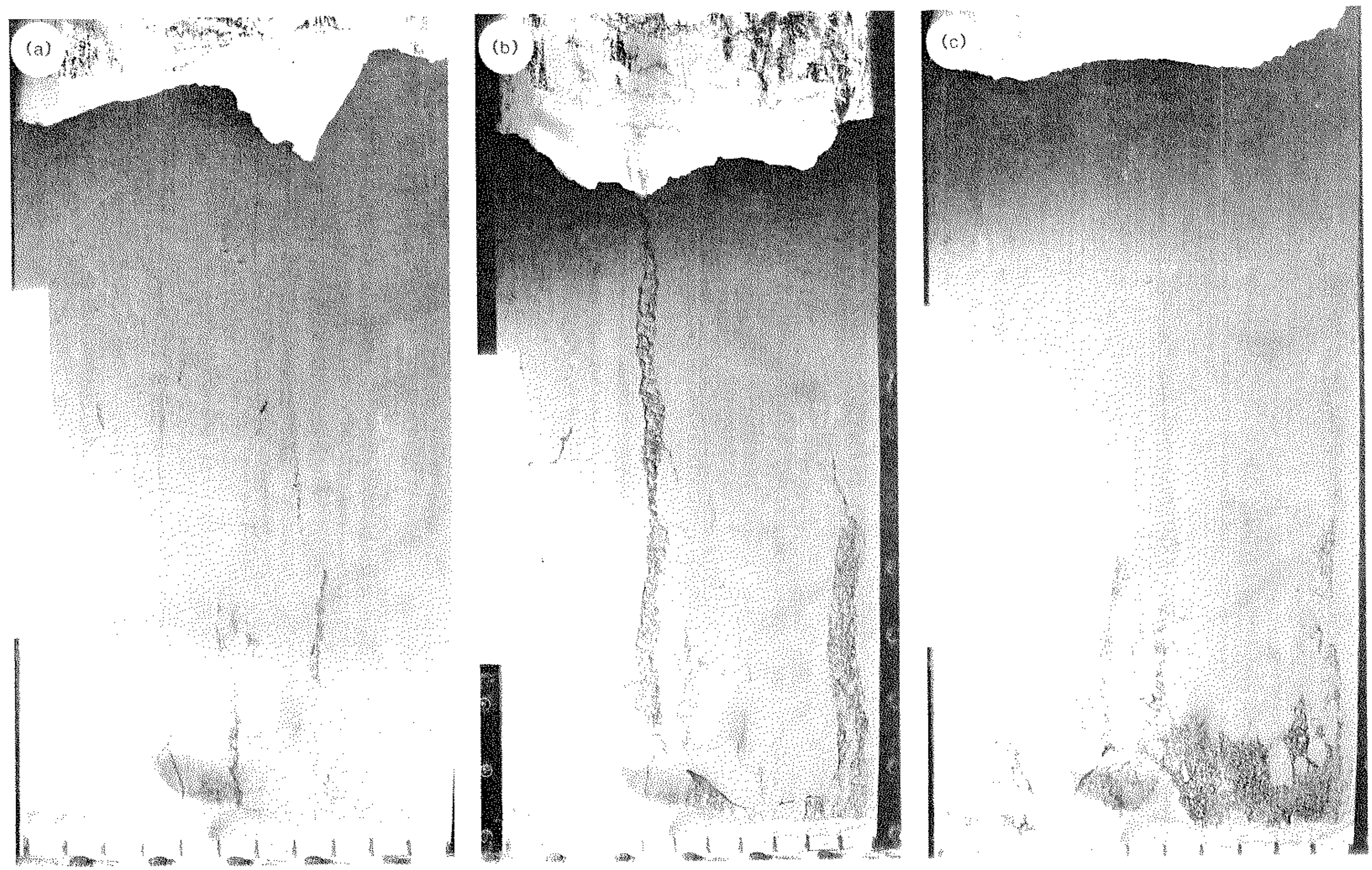
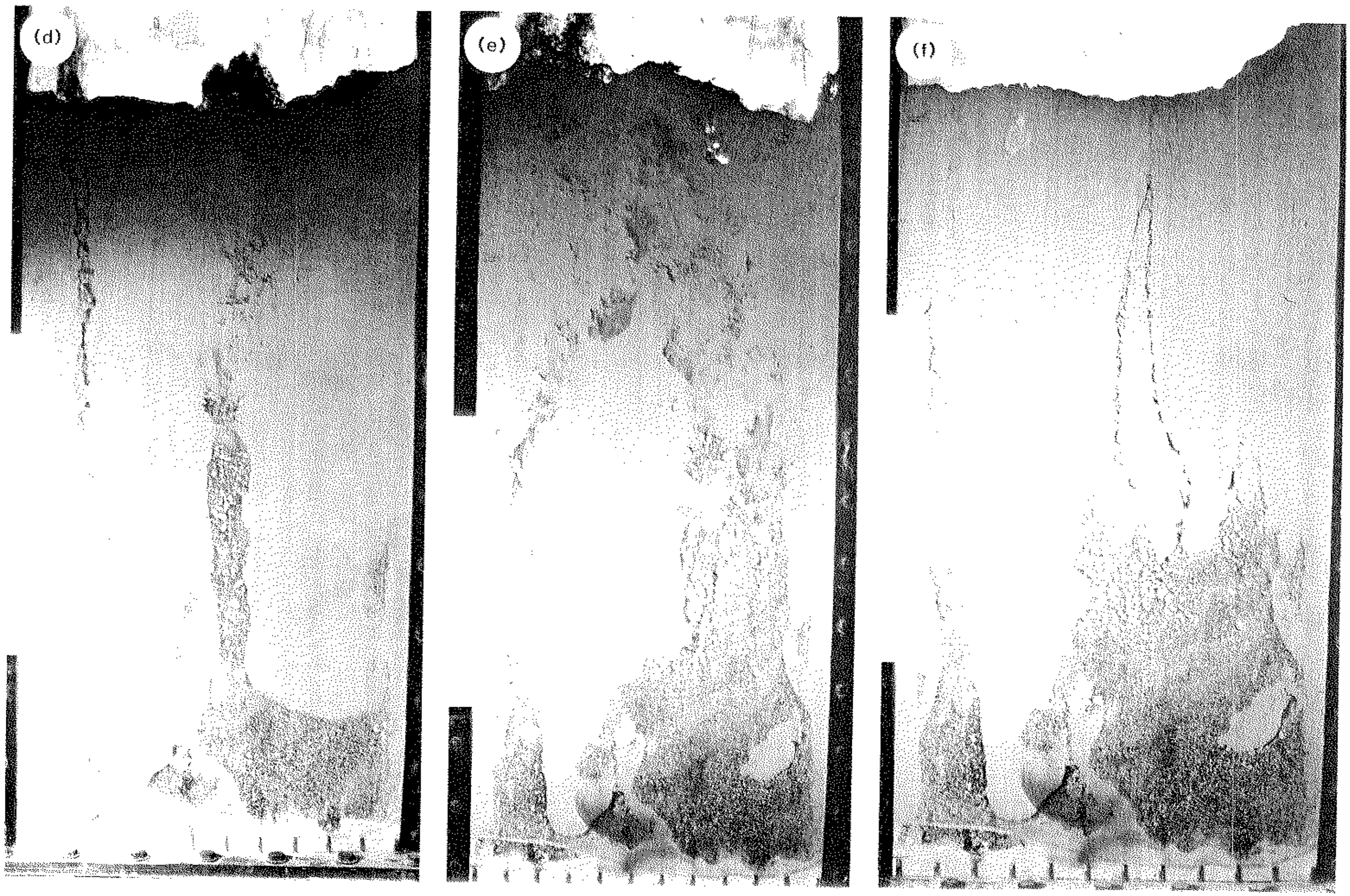



**Figure 7.7** A sequence of photographs taken through the side wall of a '2-D' fluidisation bed to show the formation of segregation features in a pyroclastic flow sample ($Md_\phi = 0.75$, $\sigma_\phi = 3.2$). Scale bar is 25 cm. (a) After a few seconds of bubbling, segregation pipes start to form. (b) As bubbling continues, the pipes become larger and propagate to the bed surface. (c) As the gas velocities are increased, the pipes are broken up by bubbling-induced circulation, and irregular segregation pods sediment towards the base of the bed. (d) With time, pipes continue to propagate back towards the

**Figure 7.7** A sequence of photographs taken through the side wall of a '2-D' fluidisation bed to show the formation of segregation features in a pyroclastic flow sample ($Md_\phi = 0.75$, $\sigma_\phi = 3.2$). Scale bar is 25 cm. (a) After a few seconds of bubbling, segregation pipes start to form. (b) As bubbling continues, the pipes become larger and propagate to the bed surface. (c) As the gas velocities are increased, the pipes are broken up by bubbling induced circulation, and irregular segregation pods sediment towards the base of the bed. (d) With time, pipes continue to propagate back towards the surface of the bed. The material ejected by the pipes is forming a fines- and pumice-rich segregation layer at the top of the bed. (e) As higher gas velocities are reached, the bed becomes distinctly layered, being richer in lithics and crystals and poorer in fines at the bottom of the bed, and richer in pumice and fines at the top. The base of this upper segregation layer is approximately at the top of the scale bar. (f) If the gas supply is slowly reduced within the upper segregation layer, then the pipes are very thin and dominantly composed of pumice, almost all of the lithics and crystals being in the lower segregation layer. (After C. J. N. Wilson 1980.)



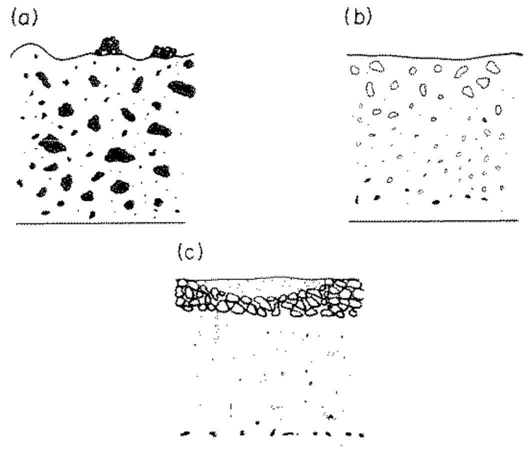



**Figure 7.8** Sketch showing the characteristics of C. J. N. Wilson's three flow types. (a) Type 1: ungraded deposit, surface ridging and block trains present, basal layer poorly developed or absent. (b) Type 2: coarse-tail grading of larger pumice towards top (reverse) and larger lithics towards base (normal), well developed basal layer, concave upper surface. (c) Type 3: strongly developed coarse-tail grading with sharply bounded pumice and lithic concentration zones, segregation bodies and pipes, and a fine-grained layer of elutriated vitric ash segregated at the top of the flow. (After C. J. N. Wilson 1980.)

which will strongly suppress turbulence in the body of a moving flow (Section 7.7).

Another important feature shown by Wilson's experiments, and indicated by the above discussion, is that fluidised pyroclastic flow samples expand much less than conventional samples do, this being due to the bypassing of the gas through segregation channels. Wilson estimated that a 100 m thick pumice flow, having fairly high gas velocities, will deflate to form an ignimbrite depositional unit which is not less than 70–85 m thick. This interpretation is also supported by '*a high tide mark*' found on the valley sides above some ignimbrites, e.g. in the Valley of Ten Thousand Smokes ignimbrite (Fig. 8.5) this mark rises only 50 m above the general level of the deposit.

Finally, we must briefly consider what the gas sources in pyroclastic flows are. These can be divided into:

(a) internal sources, being gases released from juvenile clasts by diffusion, breakage and attrition, and

(b) external sources, which include gas trapped during initial flow formation, air incorporated at the front of the moving flow, gases released by the combustion of vegetation and steam from heated surface water or ground water.

The relative importance and effects of these different gas sources are fully discussed by C. J. N. Wilson (1980).

The major part of the gas in pumice flows is provided by emission from juvenile fragments and by entrapment of air, both during eruptive column collapse and by engulfment at the flow-front. Sparks (1978b) modelled the diffusional loss of residual gases from juvenile ash particles during flow, and concluded that gas production rates are generally sufficient to fluidise fine and medium ash-sized particles in medium- and large-volume pumice flows. Large-volume (thick) pumice flows are likely to be substantially fluidised, and this may be important in determining their mobility. Increasing levels of fluidisation are predicted in larger flows and in flows with higher initial gas contents in the original magma. McTaggart (1960) suggested that the mobility of pyroclastic flows in general was due to the expansion of entrapped and heated air causing fluidisation of the flow. However, despite their high temperature, historic examples of small pyroclastic flows were no more mobile than cold rock avalanches (Fig. 7.2), which therefore throws some doubt on the importance of entrapped heated air.

## 7.3 Pyroclastic flow units and grading

As previously indicated (Ch. 5), pyroclastic flow deposits are usually composed of a number of flow units. Sparks *et al.* (1973) first proposed a layering scheme for pyroclastic flow units. The separate layers were interpreted as reflecting different depositional regimes within a pyroclastic flow (Fig. 5.13). Layer 1, the lowest layer, was thought to be the deposit of a dilute pyroclastic surge which moved in advance of the pyroclastic flow. Layer 2 was the deposit of the pyroclastic flow proper, and layer 3 was the deposit of the overriding ash cloud.

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In this section we will examine layer 2. This layer forms the main portion of most flow units, and is now thought to be deposited by the 'body' of a pyroclastic flow. Some deposits now correlated with layer 1 are thought to be sedimented in or from the flow 'head', rather than from a preceding surge, and we describe these and the *form* of a pyroclastic flow in Section 7.5. Most of the following discussion is concerned with the deposits of pumice flows. The features of layer 2 to be discussed are:

thickness
basal layers
vertical grading
gas segregation structures
lateral grading
compositionally zoned pumice flow units

### 7.3.1 THICKNESS

Flow units may vary in thickness from <0.1 m to >100 m where ponded in depressions. As an example, to show the variation within the deposits of the same eruption, the thicknesses of flow units forming the Minoan ignimbrite on Santorini are shown in Figure 7.9a.

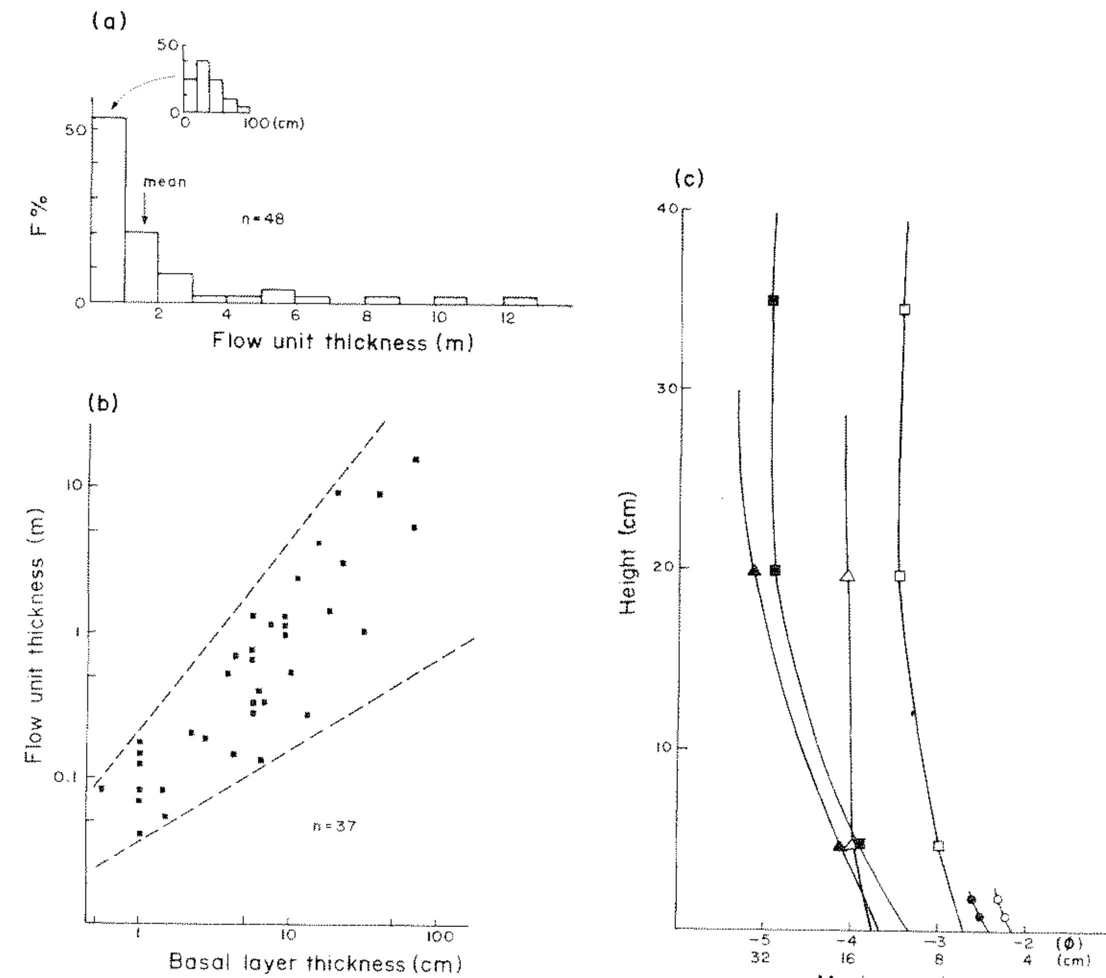


**Figure 7.9** Some data on flow units of the Minoan ignimbrite, Santorini, Greece. (a) Thickness frequency histogram of flow units. (b) Relation of thickness of the basal layer (layer 2a) to the thickness of its flow unit. (c) Plot of maximum clast size with height for three basal layers (closed symbols, lithics; open symbols, pumice).

### 7.3.2 BASAL LAYERS

Many ignimbrite flow units have a fine-grained basal layer (layer 2a) separating them from the former ground surface (Fig. 7.10), and this is usually still present even along almost vertical valley wall contacts. The thickness of basal layers varies from a few centimetres to a metre, but there is usually only a poor correlation between their thickness and the thickness of flow units within an ignimbrite (e.g. Fig. 7.9b). The basal layer differs from the main portion of the flow unit (layer 2b), in that it lacks the largest pumice and lithic clasts. Cumulative grainsize curves of the basal layer and layer 2b therefore converge towards the finer-grained size classes (Fig. 7.11). Basal layers often show a reverse grading of both larger pumice and lithic clasts (Figs 7.9c & 10), and these are often aligned parallel to the ground surface, sometimes producing a faint stratification.

All of these features suggest that the basal layer is an integral part of a flow unit and forms in response to boundary layer effects between the moving pyroclastic flow and the ground surface (Sparks 1976). The finer grainsize and reverse grading of clasts is generally attributed to grain dispersive pressures due to particle collisions acting away from the flow bottom (Sparks 1976), which is a zone of high shear. When particles of mixed grainsize are sheared together, the larger grains should drift away from the zone of maximum shear, i.e. away from the base and side of the flow in a valley (cf. debris flows; Chs 2 & 10). The alignment of clasts also suggests that the basal layer undergoes high shear stresses. Basal layers often cannot be distinguished in block- and ash-flow deposits, scoria-flow deposits and ignimbrite-flow units (see discussion below).

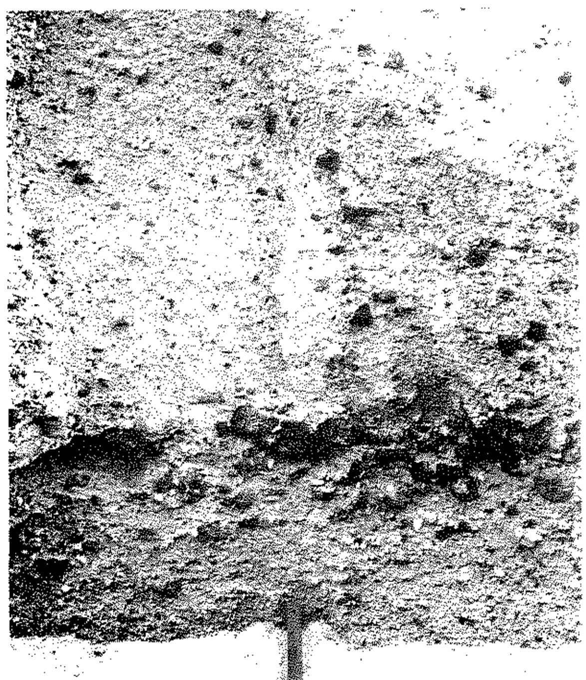

**Figure 7.10** Bottom of lowermost flow unit of the Minoan ignimbrite, Santorini, Greece, with basal layer below well defined lithic concentration zone. Flow unit deposited on top of white Minoan mud flow. Scale 20 cm.

### 7.3.3 VERTICAL GRADING

The main portion of an ignimbrite flow unit (layer 2b) may show grading of the larger clasts (Ch. 5). The 'textbook' diagram shows reverse grading of larger pumice clasts and normal grading of larger, dense lithic clasts, forming well defined concentration zones towards, respectively, the tops and bottoms of flow units (Figs 5.14c, 7.10 & 11). Flow units with reverse grading of lithics are common, but normal grading of pumice is more rarely found, and an absence of grading is also common.

Grading processes affect only the coarse part of the grainsize distribution, and this type is called coarse-tail grading (cf. distribution grading found in *classical* turbidites, Ch. 10; Allen 1982). Cumulative frequency grainsize curves of samples collected within the same flow unit converge towards the finer-grained fractions (Fig. 7.11), indicating that this part of the distribution or 'matrix' essentially remains homogeneous throughout the flow unit. Grading is therefore a function of grainsize, and is controlled by the hindered settling velocities of particles. Only large clasts have sufficiently high

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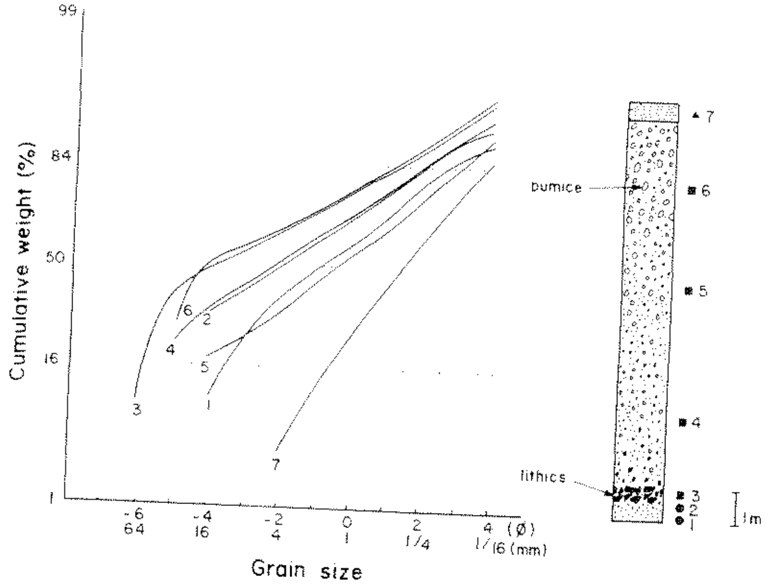


**Figure 7.11** Cumulative frequency–grainsize curves for samples of the thickest flow unit of the Minoan ignimbrite. Closed circles, samples from the basal layer; closed squares, samples from layer 2b; closed triangle is a co-ignimbrite ash-fall deposit (layer 3).

settling velocities, and therefore coarse-tail grading and poor sorting result.

The example shown in Figure 7.11 is characterised by reverse grading of low density pumice and normal grading of high density lithics. This type of gravitational (or buoyancy) grading must result from the density contrast between the larger clasts and the matrix. The matrix is invariably denser than the pumice clasts (Table 7.2), and it is thought that the flotation of the larger, lower density pumice clasts causes the reverse grading (density of pumice decreases with size). This implies that the matrix *could not have been greatly expanded* in the moving

**Table 7.2** Comparison of the densities of pumice and matrix of four flow units of the Minoan ignimbrite.

| Flow unit | Pumice density (g cm$^{-3}$) | Matrix* density (g cm$^{-3}$) |
|---|---|---|
| a | 0.58 | 1.05 |
| b | 0.55 | 1.08 |
| c | 0.64 | 1.21 |
| d | 0.54 | 1.00 |

* Matrix is <2 mm size classes.

None of these flow units showed normal grading of pumice, which indicates that the pyroclastic flows could not have been greatly expanded. In all cases the height of the moving flows could at most have only been about twice the present flow unit thickness. Flow unit a is discussed in text and shown in Figure 7.11.

pyroclastic flow, otherwise the density contrast would have been lost, and therefore the matrix density is an upper limit for the density of the moving flow. For the example in Figure 7.11 the density of the flow must have been >0.6 to <1.05 g cm$^{-3}$ (flow unit a, Table 7.2). Lithics, because of their high density (~2.5 g cm$^{-3}$) sink, and in this example the largest and heaviest clasts have formed a distinct lithic concentration zone at the base of layer 2b.

The type of grading observed within a flow unit places important controls on the properties of the moving pyroclastic flow. Normal grading of pumice may be due to more-expanded flows in which the pumice density is greater than the matrix. Reverse grading of lithics (Fig. 7.12) suggests that flows were only marginally expanded. In such flows high shear-strain rates will be imposed through their thickness and shear-induced grading of the larger clasts will result. In such flow units it is often hard to detect a separate layer 2a, because the whole flow was effectively being controlled by boundary layer effects.

Reverse grading of larger and denser clasts is similar to that commonly found in block and ash flow, and some scoria flow deposits (Figs 5.14a, & 15a & b), and again in these deposits a layer 2a is often inconspicuous or poorly developed. Nairn

**Figure 7.12** Flow unit showing reverse grading of larger lithics in the Minoan ignimbrite, Santorini, Greece. Slight erosional bench towards bottom of photograph marks finer-grained base of flow unit, but a discrete basal layer is not obvious. Scale 20 cm.

and Self (1978) stressed the importance of grain-flow (Lowe 1976; Ch. 10) and grain dispersive forces producing reverse grading in the Ngauruhoe 1975 scoria-flow deposits. However, not all reverse grading in scoria flow deposits is shear-induced, and in some deposits reverse grading of scoria clasts is controlled by gravitational grading (Fig. 5.14d). Thus, there are two mechanisms producing coarse-tail grading (within layer 2b) in pyroclastic flows in general:

(a) gravitational or buoyancy-induced grading and
(b) shear-induced grading.

These both occur in a spectrum of pumice flow types (sometimes in the same ignimbrite), and which mechanism operates is controlled by the degree of expansion of the flow.

The degree of expansion of the flow is related to the amount of fluidisation or the vertical gas flow rate upwards through the flow, and this will control grading processes. In the classification scheme of Wilson (Table 7.1, Fig. 7.8) buoyancy-induced grading is found in expanded type 2 flows, and is well developed in segregating type 3 flows. Shear-induced grading is found in unexpanded type 1 flows.

The gas flow processes involved in fluidisation produce their own grading and sorting in the finer-grained fractions. A method has been developed to analyse this type of gas grading, and the details are given in J. V. Wright and Walker (1981) and J. V. Wright (1981). The method basically assumes that high expanded flows will lose a lot of fine-grained ash (into the overriding turbulent cloud), while less-expanded flows will lose relatively smaller amounts. A qualitative assessment can be made in the field because, if the magma were porphyritic, highly expanded flows should have lost a high proportion of fine glass, and therefore the matrix should be very crystal-enriched compared with the proportion of crystals in large pumice clasts or fiamme.

The corollary is that flow units showing gravitational grading should show evidence of gas grading.

## 7.3.4 GAS SEGREGATION STRUCTURES

Gas segregation pipes in pyroclastic flow deposits were recognised by G. P. L. Walker (1971, 1972), who described them as 'fossil fumaroles'. He demonstrated that they were enriched in heavy components (crystals and lithics), and suggested that they formed by gas streaming through the ash matrix of a flow on deposition (Fig. 7.13). Earlier, H. Williams (1942) had described fumarole pipes in the Crater Lake succession. Since Walker's studies, they have been commonly found in ignimbrite flow units, and also in some denser-clast types of pyroclastic flow deposit (Figs 5.14, 15c, d & 16f). Pipes have also been found in ancient pyroclastic

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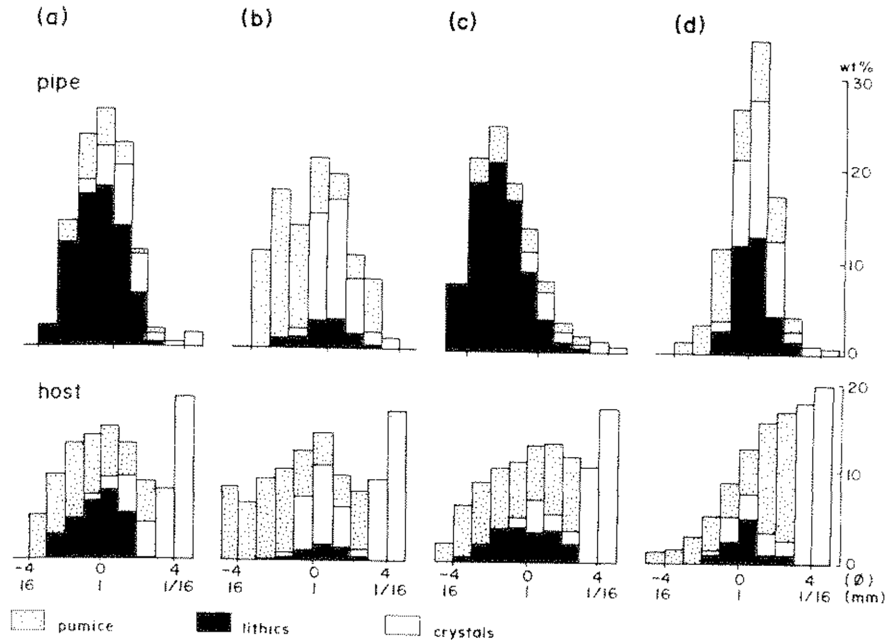


**Figure 7.13** Grainsize histograms of samples from gas segregation pipes and below the host ignimbrite which contains them. (a) Vesuvius AD 79. (b) Terceira, Azores. (c) Lake Atitlan, Guatemala. (d) Campanian Tuff, Italy. (After G. P. L. Walker 1971.)

flow deposits, where they are especially important for distinguishing pyroclastic flow deposits from epiclastic debris-flow deposits (e.g. Duyverman & Roobol 1981). Water escape pipes which may appear very similar have occasionally been found in coarse-grained sedimentary debris flow deposits (Postma 1983), including the deposits of the March 1982 Mt St Helens debris-flow event (C. J. N. Wilson *pers. comm.*).

Gas segregation structures are generally up to about 50 cm in length and a few centimetres wide, but quite often they are only several centimetres long and a few grain diameters wide. They are characteristically depleted in fines and enriched in crystals and lithics. Much larger segregation pipes >2 m long occur in coarse, near-vent ignimbrite facies (Ch. 8). In cross section they may have quite irregular shapes, and are not necessarily true pipes; pod-like structures may also be found. Segregation pipes are often more common towards the upper parts of flow units; however, they may also be found at the bases (Figs 7.14, 5.14 & 15c). These variations are thought to reflect the effects of different gas sources.

As described in Section 7.2, segregation pipes and pods can be produced during fluidisation

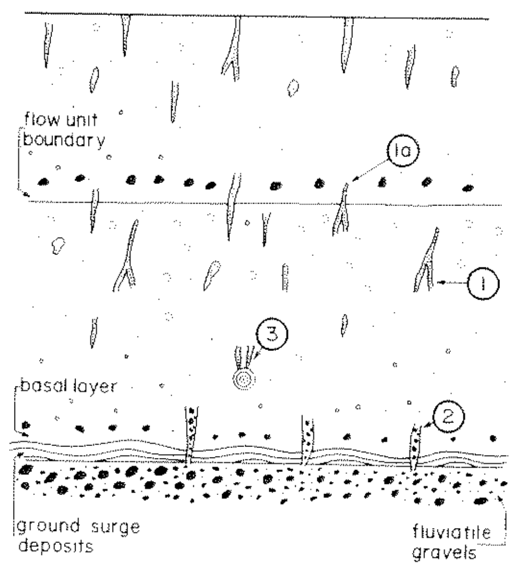


**Figure 7.14** Occurrences of gas segregation structures in pyroclastic flow deposits. 1. Pipes and pods generated initially or formed entirely by intraflow gas sources during emplacement; 1a. formed by continued post-emplacement gas flow. 2. formed from heated ground water and incorporating fluviatile pebbles; 3. formed above burnt vegetation and logs.

experiments (Fig. 7.7). Such structures are very resistant to mechanical mixing, and once formed were found impossible to destroy under laboratory conditions: they have a high yield strength because $U/U_{mp}$ is low (C. J. N. Wilson 1980). This implies segregation pipes in pyroclastic flow deposits need not always be a secondary or post-emplacement feature, but may be a primary gas-flow feature surviving from the moving flow.

It is the intraflow gas sources that are potentially the most important 'lubricating agent' and agent for the support of clasts in moving pyroclastic flows, especially pumice flows. Gas released from juvenile

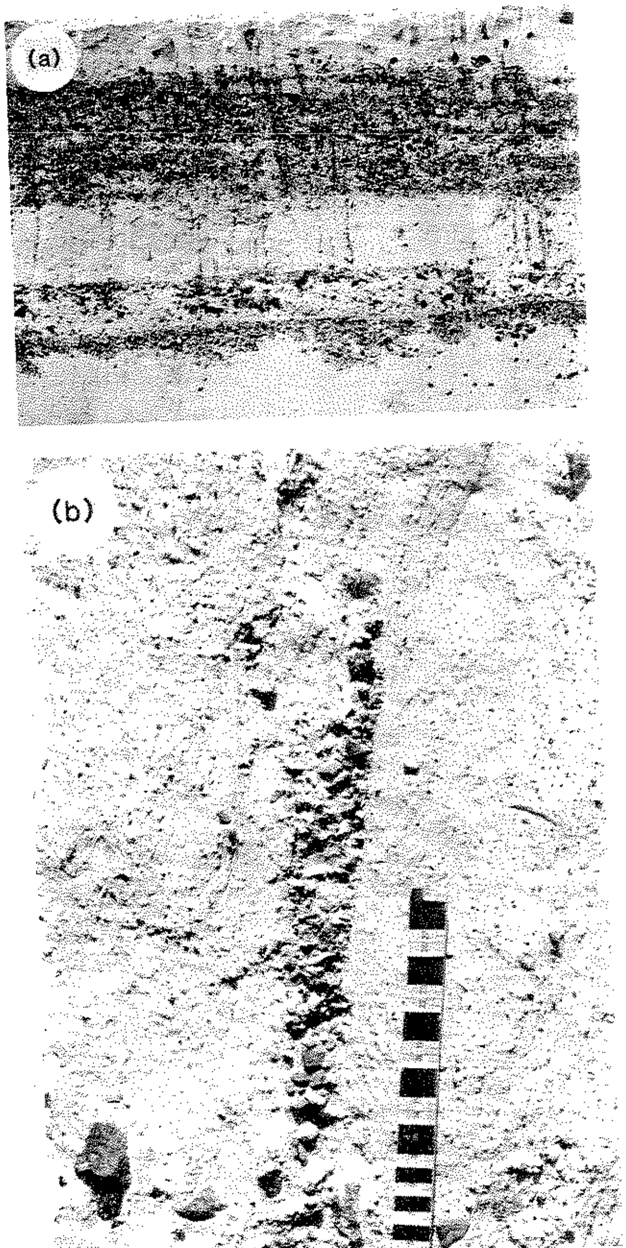


**Figure 7.15** Gas segregation pipes in flow units of the Minoan ignimbrite, Santorini, Greece, which are interbedded with coarse-grained flash flood deposits generated during the eruption.


**Figure 7.16** Remnants of a carbonised log (by lens cap) with irregular segregation pipes rising off it. This is in the distal part of a mid-Pleistocene ignimbrite in New Zealand. (Photograph by C. J. N. Wilson.)

clasts by diffusion, and breakage and attrition, should increase systematically with height through a flow, and would therefore appear to control the formation of pipes found concentrated towards the tops of flow units. Pipes of this type are now suspected to be generated in the moving flow, although once established they may also act to concentrate post-emplacement gas flow (e.g. they may cut a later flow unit: Fig. 7.14).

Pipes found at the bottoms of flow units are perhaps generally derived from external sources, and are considered to be post-emplacement (e.g. they cut basal layers; Fig. 7.14). Good examples of this type of pipe are found in the Minoan ignimbrite, where flow units are interbedded with torrent deposits and heated ground water was the gas source (Bond & Sparks 1976; Fig. 7.15). Segregation pipes can also be found above burnt vegetation and logs (Figs 7.14 & 16).

Small segregation pipes have also been recorded from pyroclastic surge deposits. Examples are found in ash-cloud surge deposits in the Bandelier Tuffs (Fisher 1979). These cut the internal lamination of the surge deposits, and were formed by post-emplacement gas flow from the underlying parent pyroclastic flow unit.

### 7.3.5 LATERAL GRADING

Small, denser-clast types of pyroclastic flows may carry the largest blocks along the full length of their run-out, and show no appreciable lateral grading (e.g. P. W. Francis *et al.* 1974, D. K. Davies *et al.* 1978a).

Many ignimbrites are known to show a decrease in the maximum size of lithic clasts with distance from source (Figs 7.17 & 18). The distance at which clasts segregate out (into a lithic concentration zone at the base of layer 2b) will be dependent on:

(a) size and density of the clasts and
(b) the density or viscosity of the 'matrix' of fluidised fines.

Using the above criteria and data on the lateral grading of large lithic clasts, Sparks (1976) estimated pumice flows to have apparent viscosities in the range 10–1000 poise. Some ignimbrites show little or no lateral grading of larger lithics, and these may

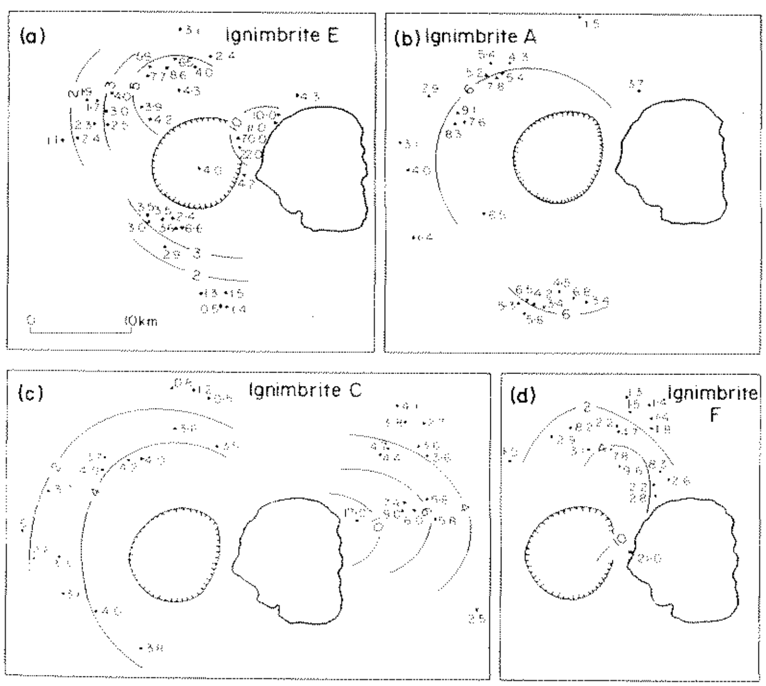


**Figure 7.17** Lateral variation in the average diameter (in cm) of the three largest lithic clasts in four of the Vulsini ignimbrites. Average size decreases away from Latera caldera (left with hachures) for ignimbrites A, E and F and Bolsena caldera (right) for ignimbrite C. (After Sparks 1975.)

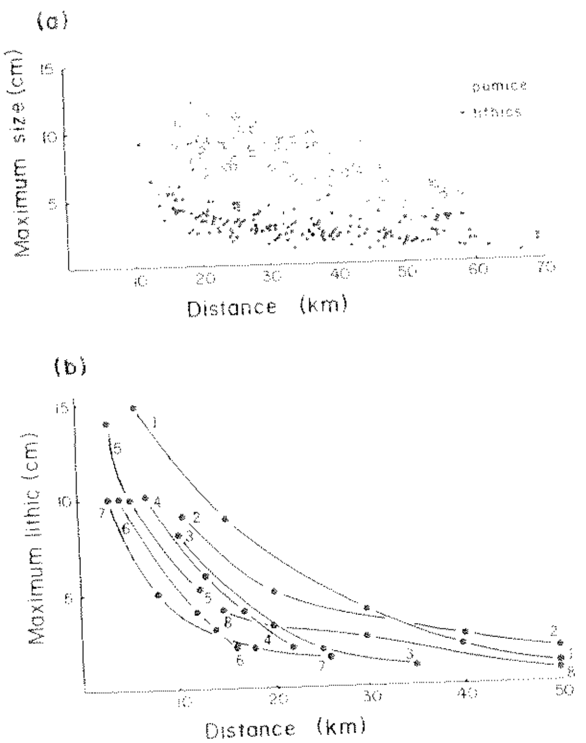


**Figure 7.18** (a) Lateral variation in average maximum clast size in the Ito pyroclastic flow, Japan (after Yokoyama 1974). (b) Relationship of the average maximum lithic clast size to distance from source for several ignimbrites: 1, Towada pumice flow deposit; 2, Ito pyroclastic flow deposit; 3, Shikotsu pyroclastic flow deposit; 4, ignimbrite C, Vulsini; 5, Lajes ignimbrite, Terceira; 6, ignimbrite F, Vulsini; 7, ignimbrite E, Vulsini; 8, Kuttyaro pyroclastic flow deposit. (After Sparks 1975.)

have had higher viscosities, e.g. the Acatlan ignimbrite, Mexico (J. V. Wright & Walker 1977, 1981).

Larger pumice clasts sometimes also show a decrease in maximum size with distance from source (e.g. Fig. 7.18). In many other cases no simple relation is found, or maximum size may even increase with distance.

### 7.3.6 COMPOSITIONALLY ZONED PUMICE FLOW UNITS

Many ignimbrite sheets show progressive vertical changes in mineral and chemical composition, representing the inverted sequence of magma erupted from zoned chambers. J. V. Wright and Walker (1977, 1981) described compositional zoning within *a single flow unit*. The upper main 30 m flow unit of the Acatlan ignimbrite passes from white rhyolite pumice, through a thin mix zone (*passage zone*), into black andesitic pumice (Figs 7.19 & 20). Other ignimbrites showing a similar change within one flow unit are also known, e.g. the Hraunfossar ignimbrite in western Iceland (Sheridan 1979) and the Kaingaroa ignimbrite, New Zealand (C. J. N. Wilson *pers. comm.*). Such compositionally zoned flow units provide another way of testing the proposed emplacement mechanisms of ignimbrites. The simultaneous emplacement of light and dark pumice layers in the same flow unit with a minimum of mixing implies a laminar flow regime for the moving pumice flow. If flow were turbulent, mixing of the two pumice types would have taken place and destroyed the compositional zoning.

For the Acatlan ignimbrite, details of the grainsize distribution suggested that laminar flow was perhaps more important early in the history of the pumice flow, and later movement of the zoned flow occurred as a semi-rigid plug moving along a sheared basal layer. The question could be asked: 'why does compositional zoning not constitute evidence of layer-by-layer deposition from a turbulent, low particle concentration pyroclastic flow, as envisaged by Fisher (1966a)?' However, the massive structureless nature of the deposit, lack of bedforms and the general lack of any vertical or lateral grainsize changes (except at vent, Ch. 8) all indicate that the flow had a high particle concentration, and deposition would have been almost instantaneous by freezing of the flow, not layer by layer.

Some care must also be taken in the interpretation of other colour zoned deposits, because such a change may be a post-depositional thermal effect.

## 7.4 Theoretical modelling of the transport of pumice flows

We have already mentioned some of the theoretical modelling by L. Wilson, Sparks and co-workers on the generation of pumice flows by collapse of plinian eruption columns (Ch. 6). Here we will

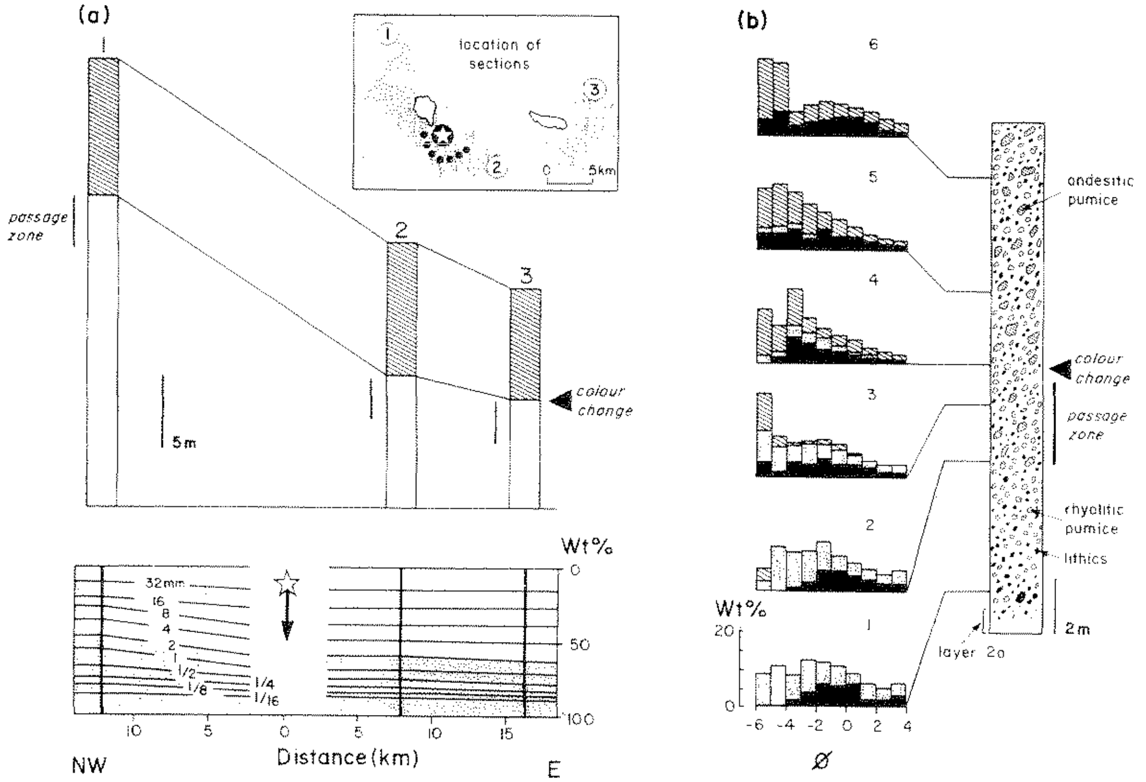


**Figure 7.19** The compositionally zoned upper main flow unit of the Acatlan ignimbrite, Mexico. (a) Profiles at three locations and average grainsize distribution. Ash content highlighted by close stipple. Star marks source position, unornamented is area in which co-ignimbrite breccias occur (Ch. 8). Inset shows lateral extent of ignimbrite and positions the three locations; again, star is source vent, and large dots enclose area in which co-ignimbrite breccias are found (Fig 8.19). (b) Details of the grainsize distribution in the profile measured at location 2. In histograms stipple is rhyolitic pumice; diagonal rule is andesitic pumice; black is lithics. (After J. V. Wright & Walker 1981.)

briefly present some results of the modelling of the movement and emplacement of pumice flows from Sparks *et al.* (1978).

The models of Sparks *et al.* (1978) indicate that the height of eruption columns, when collapsing above the vent, is between 0.6 and 9 km for the range of values of the controlling parameters considered (gas velocity, water content and vent radius) (see Ch. 8). They then postulate that when eruption column collapse occurs, the mixture flows away from the volcano as a density current with high velocity. Initial velocities of the flows range from 60 to 300 m $s^{-1}$ (Fig. 7.21). Flows from large eruptions may still have velocities of 100 m $s^{-1}$ at distances of 50 km from the source. The models show that initially the mixture is likely to be a highly turbulent, rapidly moving density current with a low particle concentration. However, a considerable proportion of the particles transported by the pumice flows are unable to be supported as a turbulent suspended load. This is because much of the grainsize distribution has terminal fall velocities well above the shearing stress velocities maintaining turbulence in a flow, even for rapidly moving flows.

The ability of turbulent flows to suspend particles is related to a parameter called the shearing stress or friction velocity, $V^*$, defined by

$$V^* = \left(\frac{cf}{2}\right)^{U} \sqrt{2} \qquad (7.9)$$

where *cf* is the drag coefficient of the ground and *U*

**Figure 7.20** Compositionally zoned upper main flow unit of the Acatlan ignimbrite showing colour contrast between the lighter, lower rhyolitic part and the darker, upper andesitic part. Note the occasional large, dark juvenile andesitic clast in the top of the lower rhyolitic part which mark the passage zone (Fig. 7.19). Height of outcrop is 4 m.

is the mean velocity. True turbulent suspension occurs when the terminal velocity, $V_t$, of a particle is less than the product of the shearing stress velocity and a constant $\beta$. Values of $\beta V^*$ were computed at the same time as the velocity distance curves shown in Figure 7.21. Figure 7.22 shows the variation of $\beta V^*$ with distance from the vent. $\beta V^*$ diminishes away from the source as the intensity of turbulence diminishes. In Figure 7.23 the terminal velocities of pumice and lithic clasts are shown in fluids of several different densities, together with shearing stress velocities for flow velocities of 10–100 m s$^{-1}$ and a drag coefficient of 0.01. These

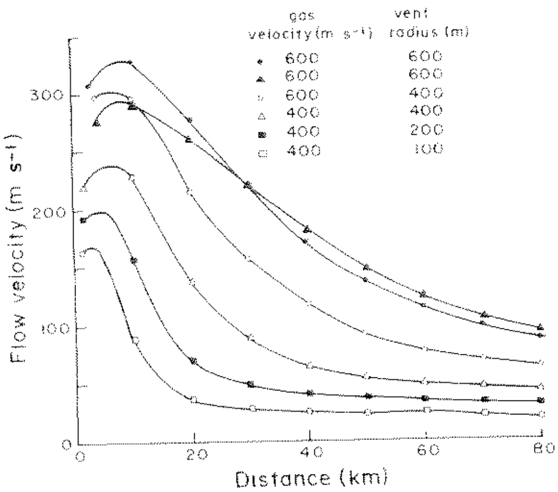


**Figure 7.21** Variation of velocity with distance from a central source in a radially expanding, low density, turbulent fluid. Six sample solutions are shown, to cover a wide range of eruption column conditions. The flow velocity is calculated for the conditions of gas velocity and vent radius indicated. All of the models are for 1% water content, except for the solid circle solution, which is for 6% water content. (After Sparks *et al.* 1978.)

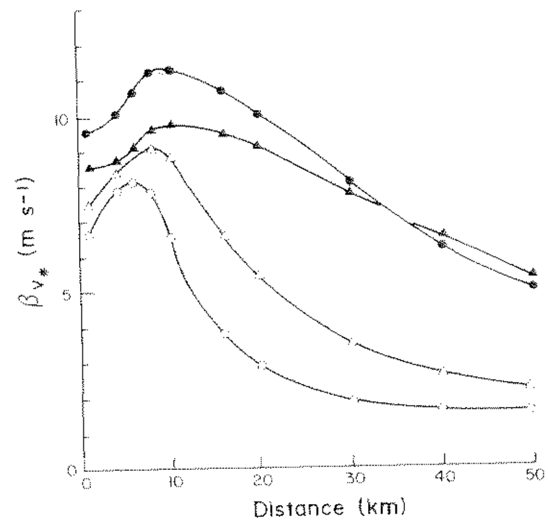


**Figure 7.22** Numerical solutions to the variation of shearing stress velocity ($V^*$) in four model flows with distance from the source. The four examples are the same as four of those in Figure 7.21, and the legend is the same. (After Sparks *et al.* 1978.)

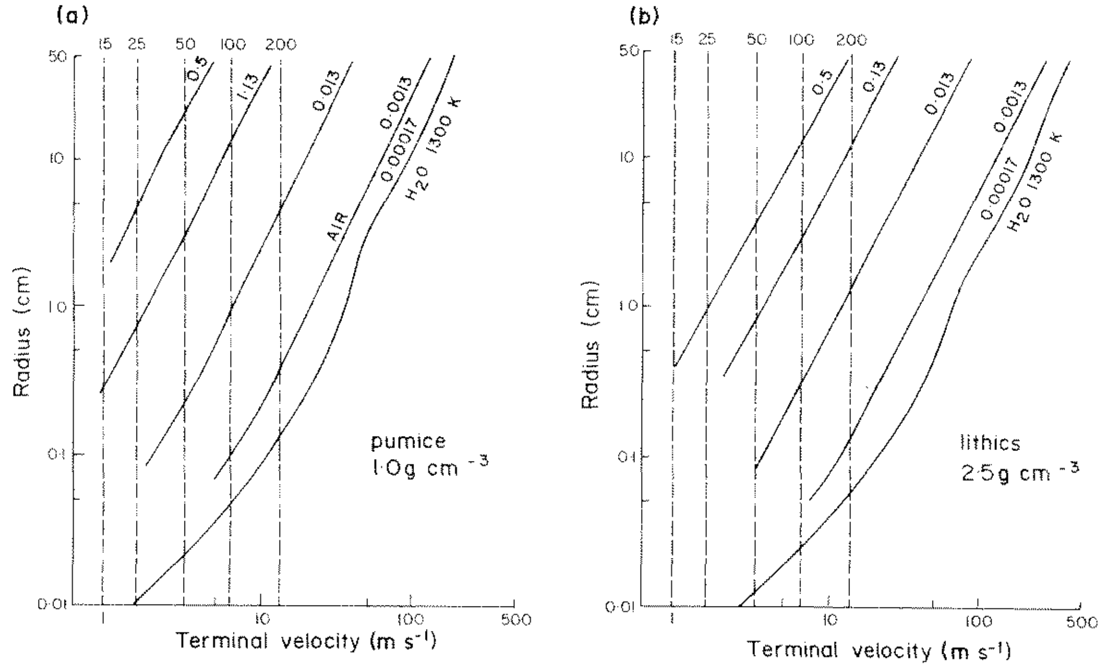


**Figure 7.23** Terminal velocity–grainsize curves are given for (a) pumice clasts and (b) lithic clasts in fluids of several different densities. The curves for the densities 0.0013 and 0.00017 g cm$^{-3}$ are for air at room temperature and steam at 1300 K, respectively. The gaseous medium in a pyroclastic flow is assumed to have a density between that of cold air ($\rho$ = 0.0013 g cm$^{-3}$) and that of hot steam. The curve for $H_2O$ at 1300 K is an exact curve using the solutions of G. P. L. Walker *et al.* (1971), whereas the other solutions plot an approximate equation (Reynolds Number of >500). The broken vertical lines represent the shearing stress velocities of flows with velocities of 15, 25, 50, 100, and 200 m s$^{-1}$ and a friction factor of 0.01. The intersection of any terminal velocity curve with any value of shearing stress velocity defines the size of the largest particle that can be suspended by a flow of that density and velocity. (After Sparks *et al.* 1978.)

show that even in fast flows only particles finer than about 1 mm can be transported in turbulent suspension and, in many cases, only particles finer than a few hundred micrometres can be suspended.

It was therefore deduced that pyroclastic flows develop a high-concentration basal zone within a few kilometres of the vent, as larger clasts settle to the base of the flow. In such a high concentration flow other mechanisms of particle support are dominant (see above) and the flow is capable of transporting lithic clasts of diameter several centimetres for tens of kilometres. The motion of the lower, dense flow dissociates itself from the upper turbulent cloud of fine ash and gas, which mixes with the atmosphere to form a convective plume.

## 7.5 Form of moving pyroclastic flows: head, body and tail deposits

The term '*form*' is used to describe the shape which a pyroclastic flow will assume when moving. C. J. N. Wilson and Walker (1982) proposed that a pyroclastic flow can be divided into a head, a body and a tail (Fig. 7.24). The head region and the body and tail region have different fluidisation states, and this controls the development of separate layers and facies within a pyroclastic flow deposit. Flows with well developed heads should produce deposits that are very different from those without heads, in which nearly all of the material is deposited from the body. It is at the head that erosion takes place, and this will be greater in flows with larger heads.

The most fluidised part of a pyroclastic flow will be the flow-head (C. J. N. Wilson 1980), into

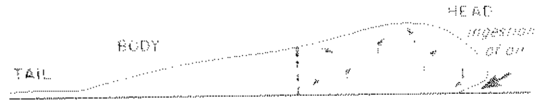


**Figure 7.24** Schematic diagram to illustrate the form of a pyroclastic flow and the division into head, body and tail.

which large quantities of air can be ingested (Fig. 7.25). By analogy with other density currents (Allen 1971, Simpson 1972), a fast-moving pyroclastic flow is likely to develop a lobe and cleft structure at its front, with air being preferentially ingested through the clefts. Ingested air expands rapidly due to the sudden temperature increase, and this causes strong fluidisation, as well as variable degrees of turbulence. Deposits sedimented out of the flow-head should therefore be more fines-depleted and enriched in crystals and lithics than those deposited by the remaining portions of the flow (Fig. 7.25).

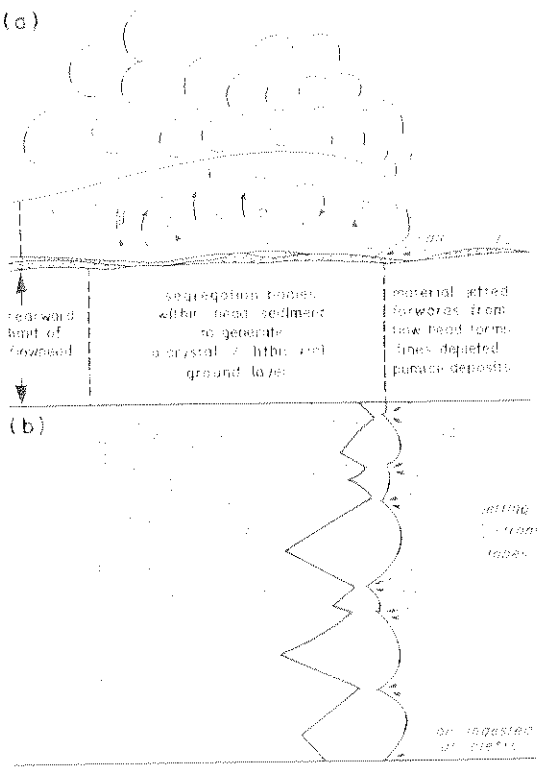


**Figure 7.25** Processes occurring in the highly fluidised flow head of a pyroclastic flow like the Taupo ignimbrite. (a) Vertical section, (b) Plan view. (After C. J. N. Wilson & Walker 1982.)

Most pyroclastic flow deposits that have been studied so far appear not to show well developed head deposits, but two exceptions are the Taupo ignimbrite, New Zealand (C. J. N. Wilson & Walker 1982, G. P. L. Walker & Wilson 1983, C. J. N. Wilson 1985; Ch. 8) and the Rabaul ignimbrite, New Britain (G. P. L. Walker *et al.* 1981c). However, it may be that a number of crystal- or lithic-rich deposits previously regarded as ground surges originated in the heads of flows (Ch. 5). G. P. L. Walker *et al.* (1981a) proposed the name ground layer for this type of head deposit. The ground layer of the Taupo ignimbrite (Figs 7.26 & 27) extends nearly to the distal limits of the ignimbrite, although there are many short breaks in its continuity. Near the vent it is a very conspicuous bed of volcanic breccia up to 3 m thick (Ch. 8). The ground layer rests on a marked erosional surface, and erosion and deposition by the flow-head must have occurred sequentially. The top of the ground layer is separated from the base of the overlying pumiceous part of the ignimbrite by another sharp erosional contact; this is almost planar, and developed by shearing when the ground layer was overridden by the rest of the flow. In the layering scheme of Sparks *et al.* (1973) the ground layer is a variant of layer 1. If true ground-surge deposits are produced by surges generated from and travelling in front of a flow (Ch. 5), then a ground layer should normally also be present (see Fig. 7.29). However, neither the Taupo nor the Rabaul ignimbrites have an accompanying ground-surge deposit.

Another type of head deposit (and variant of layer 1) has also been recognised. These, in contrast to ground layers which are concentrated in heavies (crystals and lithics), are fines-depleted pumice deposits. In the Taupo ignimbrite they form a laterally discontinuous unit below the ground layer, and are separated from it by an erosive contact, described above. Fines-depleted pumice deposits are the lowest deposit of the pyroclastic flow. Thickness varies greatly, from a few centimetres to several metres, being more common and thicker in topographic depressions. This layer is thought to represent material jetted out ahead of the flow-head (Fig. 7.25). Rapid, violent expansion of air as it is

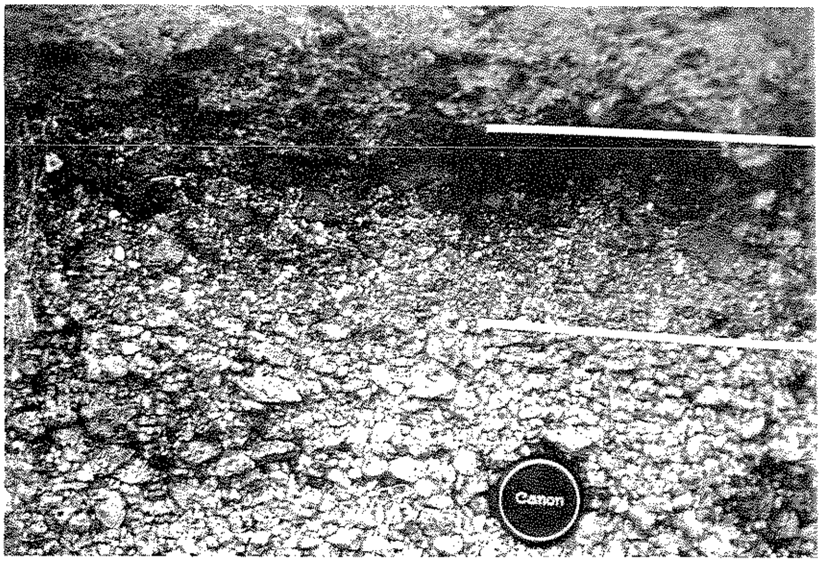


**Figure 7.26** The ground layer of the Taupo ignimbrite, New Zealand, at a location about 15 km from source (Ch. 8). The ground layer is the dark lithic rich horizon (between lines) which overlies the Taupo ultraplinian pumice-fall deposit. (Photograph by C. J. N Wilson.)

ingested into the moving flow causes portions of the pumiceous head to burst continually and to be jetted in advance of the flow proper. Large amounts of fines can be lost by this process, and in the Taupo deposit this gave rise to the distinctive fines-depleted ignimbrite or FDI (G. P. L. Walker *et al.* 1980g; Figs 7.27 & 28). At Taupo, excessive loss of fines was attributed to a very high throughput of gas, resulting partly from the ingestion of large amounts of vegetation and partly from effects of ground surface roughness on a very quickly moving pyroclastic flow, so promoting air ingestion. Data

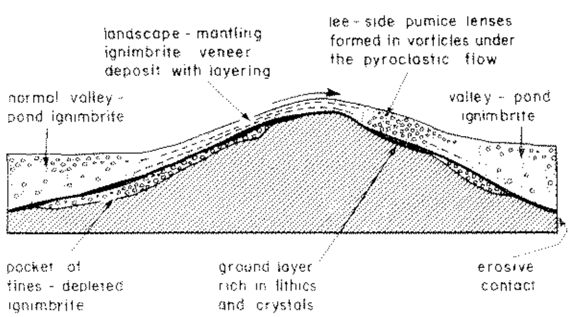


**Figure 7.27** Schematic cross section showing the facies of the Taupo ignimbrite, which are thought to have been deposited by the flow head (ground layer and fines-depleted ignimbrite), body (normal valley-pond ignimbrite) and tail (ignimbrite veneer deposit) of the pyroclastic flow. (After G. P. L. Walker 1981d.)

on the height climbed by the flow suggest that the flow velocity probably exceeded 250–300 m $s^{-1}$ near the vent (Wilson 1985). These are the highest velocities yet inferred from field observations for a pyroclastic flow, although theory suggests that they may not be unusual (Fig. 7.21). Carbonised vegetation is found at all levels in the FDI, indicating a thorough mixing by turbulence and supporting the theory of ingestion and volatilisation of large volumes of vegetation. In other fines-depleted ignimbrites, turbulence, induced locally by surface roughness, may have been the most important cause of fines loss, e.g. on St Lucia (Fig. 13.35) where pumice flows travelled down narrow, winding and heavily vegetated valleys (J. V. Wright *et al.* 1984).

The presence of jetted, as opposed to surge, deposits at the base of the Taupo ignimbrite may be related to the high velocity of the flow. Only packets of material with a high solids concentration could burst forward with sufficient velocity to be deposited before they were caught up by the flow proper. It is only where the parent flow proceeds more slowly that dilute surges have sufficient velocity to move ahead and produce a ground surge deposit (e.g. as at Santiaguito in 1973, Ch. 5). C. J. N. Wilson and Walker (1982) proposed a velocity-dependent hierarchy of conditions occur-

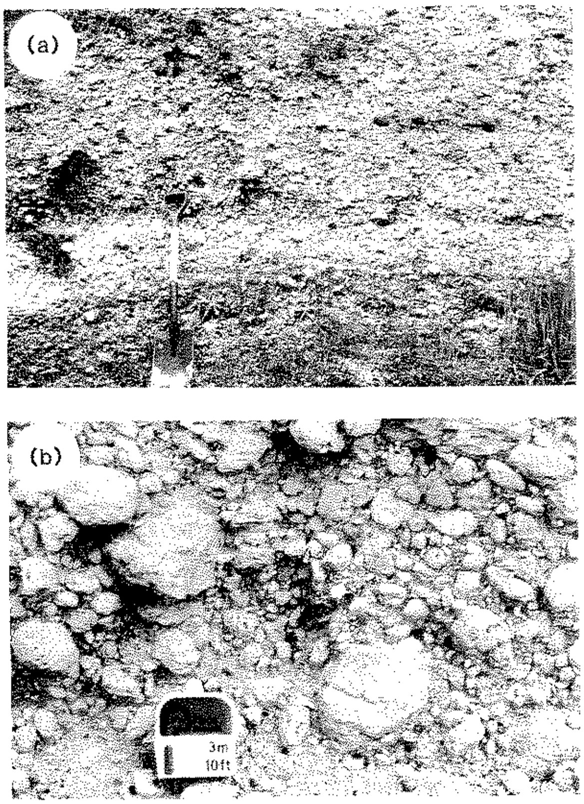


**Figure 7.28** Fines-depleted pumice deposit formed by jetting from the flow head of the Taupo ignimbrite, New Zealand. Note the clast-supported textures

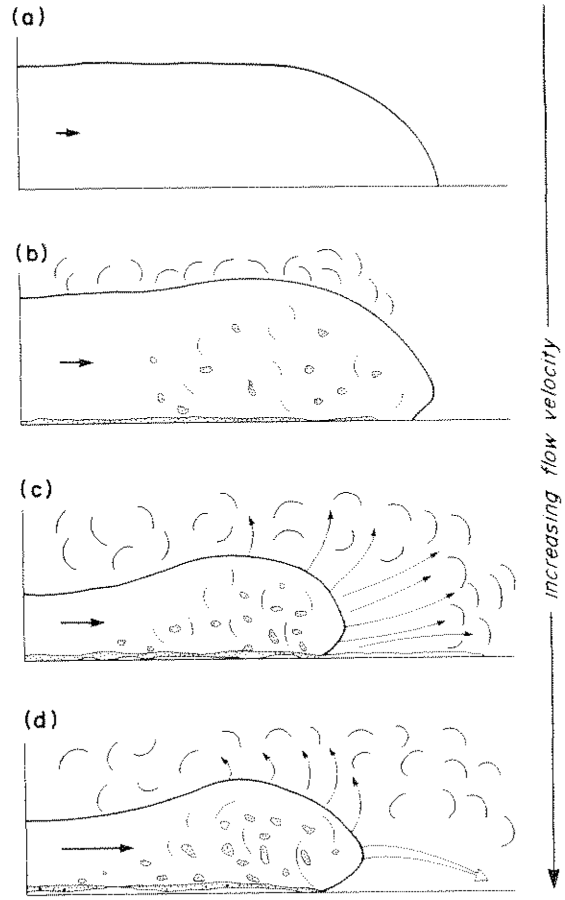


**Figure 7.29** Hierarchy of conditions found at the fronts of pyroclastic flows and the formation of various layer 1 facies. (a) The flow is so slow that no significant air ingestion occurs. No layer 1 deposits are generated. (b) Minor amounts of air ingestion cause fluidisation and segregation within the head, causing the formation of a ground layer. (c) Moderate amounts of air ingestion cause dilute surges to be generated from the front of the flow, producing ground-surge deposits. Segregation within the head forms a ground layer. (d) Strong air ingestion causes the *en masse* jetting of material from the flow-front, forming jetted fines-depleted pumice deposits. Segregation within the head forms a ground layer. The estimated flow-front propagation velocities in the above groups are very approximately: (a) 0–10 m s$^{-1}$, (b) 10–30 m s$^{-1}$; (c) 30–80 m s$^{-1}$; (d) 80–200 m s$^{-1}$. At extremely high velocities (>200 m s$^{-1}$) the situation at the flow-front is uncertain. Evidence from the near-source outcrops of the Taupo ignimbrite implies that jetted deposits are not formed, and only a ground layer containing minor amounts of fine material is generated. What form the flow-front takes is not known. (After C. J. N. Wilson & Walker 1982.)

ring at the flow-head of a pyroclastic flow, leading to the formation of the various layer 1 facies (Fig. 7.29).

The body and tail of a pyroclastic flow give rise to layer 2 deposits. In most examples the body–tail region must have suddenly stopped, with a conventional flow unit (Section 7.3) coming to rest. Layer 2 deposits in the Taupo ignimbrite comprise two very different facies: **valley pond ignimbrite**, or **VPI** (Plate 7), and **ignimbrite veneer deposit**, or **IVD** (G. P. L. Walker *et al.* 1980b, 1981b, C. J. N. Wilson & Walker 1982, G. P. L. Walker & Wilson 1983, G. P. L. Walker 1983). VPI shows features that are typical of a normal pyroclastic flow unit. The IVD is very different, and mantles the landscape. It is stratified in localities near to the vent, and occasionally shows bedforms (Fig. 7.30), and for these reasons there has been debate whether this

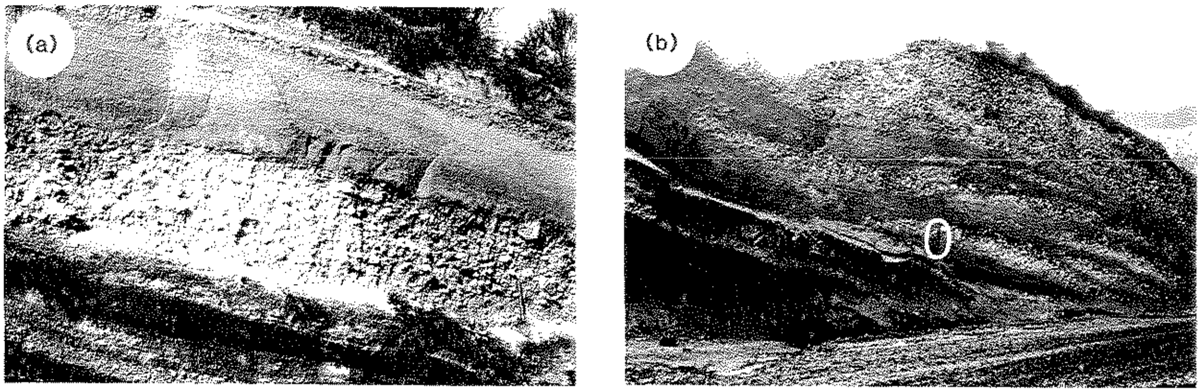


**Figure 7.30** The ignimbrite veneer deposit of the Taupo ignimbrite. (a) Finely stratified IVD overlying coarse FDI; locality is 16 km from source. (b) Coarse pumice lenses in IVD on lee-side of topographic obstacle; flow direction left to right, locality is approximately 15 km from source. Note shovel for scale. (Photographs by C. J. N. Wilson.)

deposit is a type of surge or not. However, the weight of the evidence suggests that it was left behind as a trail-marker by the tail of the flow as it moved over topography. C. J. N. Wilson and Walker (1982) suggested that the tail consisted of the lower parts of the flow which, because of their proximity to the ground surface and their lower fluidisation state, were moving less rapidly than the bulk of the flow, and hence were left behind. Grainsize stratification is thought to represent the passage of waves of material in a continuous high concentration flow, each wave depositing a layer. These layers are developed out to about 40 km from the vent, and the number decreases outwards.

Other bedforms are found on the lee-side of obstacles, where the flow jumped the ground surface. Lee-side lenses of pumice developed in turbulent vortices under the fast-moving flow, and sometimes large prograding foresets are developed (Fig. 7.30b).

IVD and VPI facies have been recognised in some other ignimbrites. They are described from the Rabaul ignimbrite (G. P. L. Walker *et al.* 1981c), and examples which were first called ash-

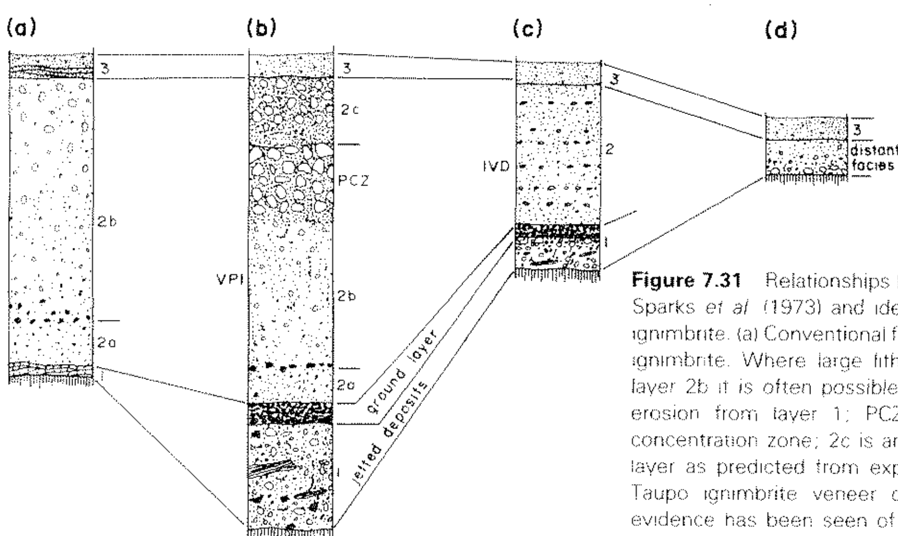


**Figure 7.31** Relationships between the layering scheme of Sparks *et al.* (1973) and idealised sections from the Taupo ignimbrite. (a) Conventional flow unit. (b) Valley-ponded Taupo ignimbrite. Where large lithics are present at the base of layer 2b it is often possible to show they were derived by erosion from layer 1; PCZ is a well developed pumice concentration zone; 2c is an upper fine-grained segregation layer as predicted from experimental studies (Fig. 7.8). (c) Taupo ignimbrite veneer deposit. (d) Distant facies. No evidence has been seen of ash-cloud surge deposits (layer 3a, Fig. 5.12) above the Taupo ignimbrite. (After C. J. N. Wilson 1985.)

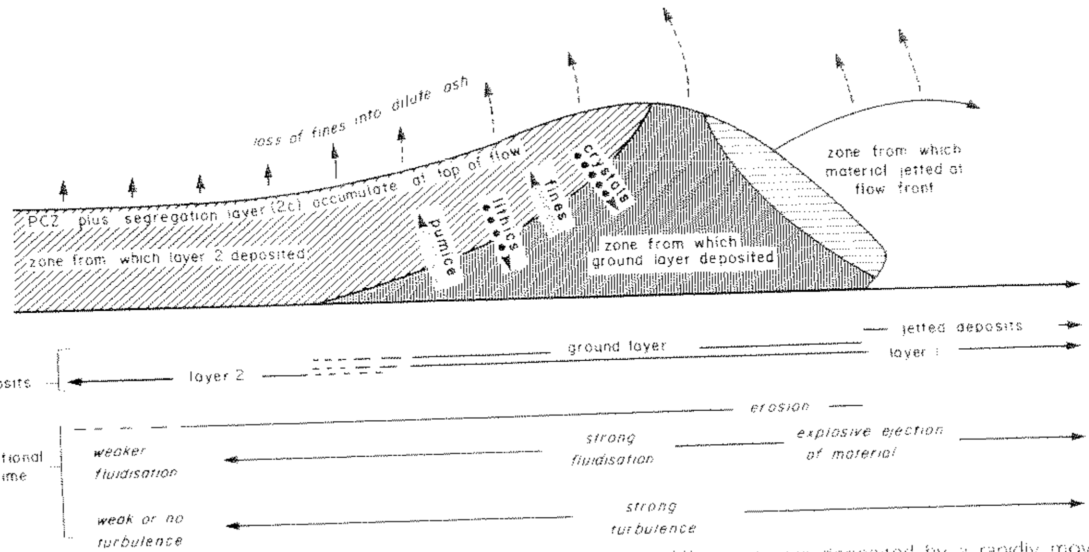


**Figure 7.32** Dynamic model summarising processes which generate the different layers deposited by a rapidly moving pyroclastic flow. (After C. J. N. Wilson 1985.)

hurricanes by Roobol and Smith (1976), occur on Martinique, erupted from Mt Pelée.

Depending on the flow velocity, and the nature of the landscape over which emplacement occurs, we can envisage a complete range of pyroclastic flow forms. Slowly moving, denser-clast flows and some pumice flows only have poorly or moderately inflated heads. These produce pyroclastic flow deposits, which in section have no, or only thin head deposits. Material transported in the flow-head is dumped, forming a steep flow front, or is pushed aside by the advancing denser body of the flow which has a much higher yield strength, so forming levées (e.g. the Mt St Helens 22 July and 7 August 1980 deposits: L. Wilson & Head 1981; Fig. 7.1b). The bulk of the flow in these cases forms a conventional layer 2 deposit. Layer 1 may include a ground surge deposit (Fig. 7.31a). Other pyroclastic flows (exemplified by the Taupo ignimbrite) which are emplaced at very high velocities have highly inflated, turbulent heads, leading to the development of prominent head deposits and a highly erosive base below a ground layer. It is envisaged that as such a flow moves, material from the flow body is laterally transferred and cycled through the more strongly fluidised front of the flow. In this region bubble-induced turbulence, due to strong fluidisation accompanying air ingestion, is important. The ground layer is then overriden by the base of the flow body, and high shear stresses result in the erosion of layer 1. These flows produce well defined body and tail deposits. The interpretation of the various facies of the Taupo ignimbrite in terms of the three-layer scheme of Sparks *et al.* (1973) is given in Figure 7.31. A dynamic model summarising their formation is shown in Figure 7.32. Such flows seem to stop because they run out of material rather than because they freeze. For this reason, the edge of the Taupo ignimbrite is not defined by a steep flow front. Instead a zone 3–5 km wide is found where layers 1 and 2 cannot be distinguished. This single layer is termed the distant facies (Fig. 7.31d), and is regarded as the deposit of the flow at the stage where air-ingestion fluidisation was affecting the entire flow, spreading a thin, strongly fluidised layer across the landscape (C. J. N. Wilson & Walker 1982, C. J. N. Wilson 1985). The flow presumably terminated when remaining material rose in the form of a buoyant convective plume adding to a co-ignimbrite air-fall ash.

In summary, we still have much to learn about pyroclastic flows and their deposits. There appears to be a very large spectrum of types, including dense block and ash flows at one extreme, to turbulent, low density, violent types at the other extreme. The latter, although only recently recognised, may be more common than is currently realised. Clearly, much more work still needs to be done on all aspects of pyroclastic flow processes and on the characteristics of their deposits.

## 7.6 Pyroclastic surges as low particle concentration flows

Pyroclastic surges are regarded as *turbulent, highly expanded, low particle concentration flows* (Ch. 5). Wohletz and Sheridan (1979) described surges as time-transient, unsteady flows of tephra that occur as a pulse or series of pulses in which the kinetic energy rapidly decays. Surges are complex *three-phase systems*, being mixtures of *solids*, *gases* and *water*. The proportions of these phases vary from one surge to another, and even during the flow of individual surges (Allen 1982), but volumetrically the proportion of solids (and thus their concentration) is subordinate to gas and liquid in most surges, and therefore much less than in pyroclastic flows (also see G. P. L. Walker 1983). In *hot, dry surges* the liquid phase is an insignificant component. In *wet, lower temperature surges* subequal proportions of all three phases are likely. In terms of analogy with epiclastic sediment transport processes, surges are therefore roughly akin to subaqueous turbidity currents (*but* see Section 7.11 for further critical discussion), whereas most pyroclastic flows are akin to high concentration, viscous regime debris flows (Ch. 10). Surges can result from eruptions of any magma type, and can arise from both magmatic and phreatomagmatic eruptions (Chs 5 & 6).

During the initial stages in pyroclastic surges, solid particles are widely dispersed within the fluid phase(s), and are essentially supported and driven by them through turbulence in the gaseous phase. The solids thus behave in a particulate fashion during transportation, and are free to sort themselves hydraulically. In viscous mass flows particle interaction and particle–fluid cohesion hinders sorting to a large degree. As the wetness of surges increases, particle freedom decreases due to adhesion processes (see below).

Surges have densities higher than the ambient atmospheric density, so their passage will be largely gravity-controlled. Generally, they follow topographic lows (J. G. Moore 1967, Waters & Fisher 1971) but, because of their largely turbulent nature, stemming from an initial explosive thrust or high gas content relative to solids (especially fines, see G. P. L. Walker 1983), or from both, they also have the ability to climb very significant topographic highs and to mantle them with a thin veneer of ash (e.g. Fig. 5.1; Nairn 1979).

Surge-forming eruptions are usually composed of multiple surges, frequently as pulses with negligible time intervals. Surge-forming eruptions may therefore lead to deposits which are stacks of surge-deposited beds. Successive surges may erode and rework to varying degrees the deposits of preceding surges.

## 7.7 Energy sources and initiation of surges

In Chapter 5 surges were subdivided into three main types:

base surges
ground surges
ash-cloud surges

Each has a different origin, and may therefore contain different proportions of the three main phases (solid, gas and liquid). This may in turn lead to surges with varying physical properties, transportational and depositional modes, and facies characteristics of their deposits.

### 7.7.1 BASE SURGES

Base surges (J. G. Moore 1967) originate from the base of a phreatomagmatic eruption column as an outward moving, ground-hugging, turbulent cloud of fluid and ash (Ch. 5, Fig. 5.17). They develop from phreatomagmatic blasts in the vent, which eject dilute mixtures of solids, gases and steam. These expand rapidly on eruption and then spread radially, or flow along directed paths as turbulent

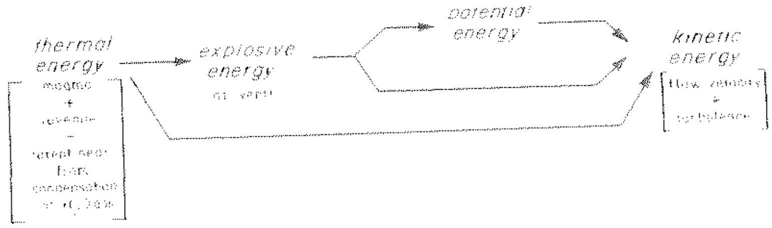


**Figure 7.33** The various energy forms involved in the initiation and flow of pyroclastic surges.

surges. Other surges originate from minor column collapse events (Ch. 5). Base surges will generally be wet, unless they are extremely hot when erupted and so contain little condensed water.

The energy for base surge initiation and flow is derived initially from the thermal energy of the rising magma (Fig. 7.33). This is translated on contact with ground water or surface water to mechanical explosive energy, as the heat from the magma is used to superheat and explosively expand the external water source (Fig. 7.33; Ch. 3). The explosive energy is translated into momentum and kinetic energy. Potential energy is also created during eruption, especially where base surges are derived from minor column collapse (Ch. 5). This potential energy, in turn, translates into kinetic energy. In addition, during at least the early stages of flow, extra thermal energy is liberated from juvenile fragments, and latent heat is released when condensation of gaseous water to liquid water droplets occurs (Fig. 7.33). The thermal energy produces turbulent convective circulation, and the large initial explosive thrust ensures high lateral transport velocities, which are responsible for turbulent flow conditions. In the Reynolds Number criterion for turbulence (Eqn 2.5, $Re = Ul\rho/\eta$), the low viscosity of the driving fluid (gas) and the high velocity fo the surge both contribute to high Reynolds Numbers. Both the convective and velocity-induced turbulence entrain and support grains (see below).

### 7.7.2 GROUND SURGES

Ground surges are associated with pyroclastic flow forming processes, and are identified where they directly underlie pyroclastic flow facies. They may have several origins, as outlined in Chapter 5:

(a) directed blasts at vent,
(b) partial collapse of maintained eruption columns and
(c) projection from the head of moving pyroclastic flows.

In the first two cases the energy for initiation and the mobility of the surge originates in much the same way as for base surges. However, those that are spawned from moving pyroclastic flows may involve one further step. Their origin may be due to the ingestion of cold air in clefts at the flow front, or

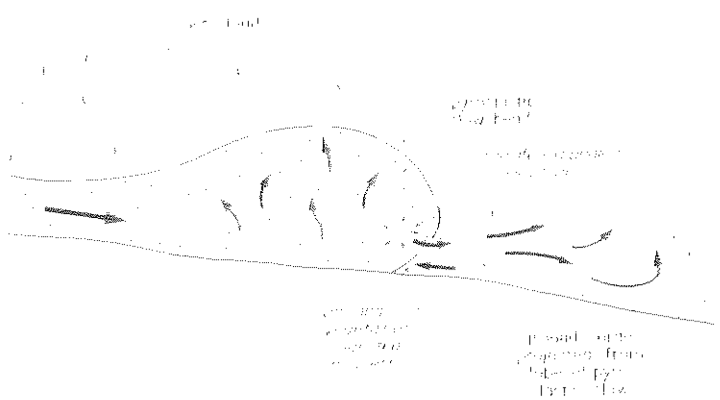

**Figure 7.34** Schematic representation of the formation of a ground surge through ingestion of air or volatilisation of vegetation in front of the head of a moving pyroclastic flow.

by erosional incorporation of excessive amounts of vegetation with its high water content (Section 7.2; Fisher 1979, C. J. N. Wilson & Walker 1982). Both of these elements would be instantly heated, and would produce rapidly expanding cells which could have enough momentum and excessive kinetic energy to be ejected forward, out of the head of the moving pyroclastic flow as a cell of highly turbulent gas and vapour with low particle concentration, which moves forward as a surge (Fig. 7.34). The surge deposit is then immediately overridden and buried under the basal facies of the following pyroclastic flow (Figs 5.13, 23a & b; Sparks & Walker 1973, C. J. N. Wilson & Walker 1982; Section 7.5).

### 7.7.3 ASH-CLOUD SURGES

Also associated with pyroclastic flows are ash-cloud surges (Ch. 5), forming from the trailing ash cloud which billows above and behind a pyroclastic flow, as gas and fine ash stream out of the head and body of the flow under the influence of fluidisation processes (Fisher 1979; Section 7.2). This loss of fine ash is called elutriation, and was discussed in Section 5.2.

The energy in this type of surge is derived entirely from the parent pyroclastic flow. The surge acquires its momentum and kinetic energy from the pyroclastic flow out of which it stems. Unlike the parent pyroclastic flow with high particle concentration, which is essentially flowing in laminar fashion (Section 7.2; Sparks 1976; but see above), the associated ash-cloud surge flows turbulently. This turbulence also has two sources – convective circulation and the overall high velocity of the surge, which contributes to high Reynolds Numbers, as discussed above. The turbulence supports the solids.

## 7.8 Transportation and grain-support processes in surges

It has been suggested above that the transportation of pyroclastic detritus in surges is dominated by turbulence in the supporting gaseous phase. This is based on direct observations of recent base surges (J. G. Moore *et al.* 1966, J. G. Moore 1967, Waters & Fisher 1971, Kienle *et al.* 1980, Self *et al.* 1980) and on the sedimentary structures observed in the deposits of all types of surges. Observed surges are dominated by turbulent billowing clouds of gas, steam and ash. On this basis, at least the peripheral parts of surges are turbulent. However, the unobservable inner, lower parts, or the 'body' of the surge, can also be considered to be turbulent by virtue of the tractional sedimentary structures such as waves (dunes), cross-stratification and horizontal lamination commonly found in their deposits (J. G. Moore 1967, Fisher & Waters 1970, Waters & Fisher 1971, Wohletz & Sheridan 1979, Allen 1982, Leys 1982; Figs 5.20, 22 & 23). All of these structures can only be produced by grain-by-grain tractional transport (rather than *en masse*; Ch. 10) implying low grain concentrations in the transporting medium. Under these circumstances, the *uplift force of turbulent fluid eddies* is the main grain-support process possible during the bulk of surge movement. Suspended grain-support will be maintained as long as the *drag force* of these uplifting turbulent eddies exceeds the settling velocities of grains at the appropriate Reynolds Number for the turbulence level (Allen 1982).

In wet surges, this simple analysis is complicated by the process of clumping or adhesion of particles in the very moist atmosphere of the moving surge (Allen 1982, Leys 1982). Clumping and adhesion refer to the aggregation of moist grains, especially fines, with moisture droplets, so increasing the *effective dynamic transported grainsize*. For continued suspension transport of these clumps, the upward component of the fluid drag force must therefore also exceed the settling velocity of the clump. The criterion for turbulent suspension has been discussed in Equation 7.9. Allen (1982) suggests that a simple criterion for suspension transport to be predominant is a Bagnold criterion:

$$W/V^* \leq 1.25 \qquad (7.10)$$

where $W$ is the particle settling velocity and $V^*$, as defined in Equation 7.9, the shear velocity of a moving flow or current.

These conditions hold for an initial highly turbulent state and for small grainsizes. However, surge deposits are known to contain fragments several centimetres or more in diameter, which were not finally transported and deposited as ballistic blocks. They must have been at least entrained, and moved by surges as well. The actual competence of surges to transport different sizes in true suspension is not clear, in spite of the discussion related to Equation 7.9. However, it is clear that during incipient sedimentation, if not before, a large proportion of the solids are saltated (bounced), rolled and dragged over the substrate as bed load under the influence of a large lateral shear velocity (Ch. 10). More of the population will experience this mode of transport as the surge loses momentum and begins to deflate.

There is therefore very likely to be a 'grain carpet' layer at the base of most surges, out of which the bulk of sedimentation will occur, as well as a more dilute trailing ash cloud. The grain-layer population is dragged along by the shear stress operating at the base of the flow. Individual grains will propel neighbours by colliding with them and transferring momentum, which constitutes a form of inertial *dispersive pressure*. It can be speculated that a well defined, voluminous grain layer would be most common in base surges. Base surges originate as a vent blast and propel a poorly sorted grain population from which a significant coarse fraction will tend to drop out quickly. Both ground surges and ash-cloud surges consist of a population that is presorted to some degree. Therefore, in a single surge some transport will take place through turbulent suspension and some through shear-induced traction and grain collisions. As the surge loses energy and deflates, more solids, particularly the coarser fractions (including clumps) will cease being transported in a suspension mode and will become entrained in a traction mode of transport. At low grain concentrations, discrete grain traction probably occurs; at higher concentrations a grain-layer flow with high degrees of particle interaction and collision, propelled by shear-stress, will develop. The change from low grain concentration traction to high concentration, shear-induced grain layer flow, conditions should be reflected, respectively, by cross-stratification and discrete lamination, and by more massive or diffusely layered beds.

Complications to this will arise according to the wetness, or moisture content, of the surge. In very dry surges, the grain transport can be viewed as occurring in an inertial regime, i.e. free from viscous interaction with other grains or with the driving fluid, as described above. As the moisture content increases, and cohesive clumps of sediment develop, transport becomes pseudo-viscous near the base of the surge. It is still not wholly viscous, because the high shear stress operating probably continually reconstitutes wet clumps and smears them out. That is, the moisture-laden, driving gaseous fluid is distinct from the muddy clumps it is propelling, since they have not totally amalgamated into a cohesive, viscous fluid. In this sense, wet surge transport is quite different from other types of particulate tractional transport or mass-flow transport because it is a *polyphase system*. It has not been mathematically modelled. Evidence of this complex phase transportation by wet surges is found in accounts of observed base surges and their deposits. Plastering of muddy ash on the upwind side of trees and buildings was documented at Taal (J. G. Moore 1967). That these surges were still turbulent is attested by the 'sandblasting' effects in abrading and stripping of bark and foliage, and in the near-blast zone, by the total destruction of vegetation (J. G. Moore 1967; Fig 5.19).

The role of fluidisation as a grain-support process and the origins of the gaseous component in surges has been touched on by several workers (Fisher 1979, Allen 1982, Leys 1982). Fluidisation in polydispersive systems is most effective where the grain concentration is high, and an abundance of fines exists to stem the free flow of fluid through the dispersion (Allen 1982, G. P. L. Walker 1983). In pyroclastic flows it is seen as a very effective agent in maintaining the mobility of the flows (Section 7.2). In surges it is less likely to do so because of the low concentration of solids right from the outset. Any fluid which is ingested will not be concentrated along discrete channels, but will mix with the ambient fluid phases in the surge. In this sense, the ingestion of air at the head of the surge may temporarily increase the internal turbulence of the surge, particularly if the air is cold and the surge is hot, so leading to expansive heating of the ingested

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air. However, there will be overall loss of heat and energy from the surge. Leys (1982), following Allen (1971, 1982), suggested that because there is only a small overhang in the profile of the head or leading edge of observed surges, it is unlikely that significant ingestion of external air into the head and at the base of the head occurs. If this is so, then ingestion and mixing, and fluidisation in the head region of surges, can be discounted as significant grain-support processes. The absence of known gas and segregation pipes originating *within* surge deposits (cf. pyroclastic flow deposits, Section 7.3.4) also supports this. This is not to say that surges do not contain gas and air, which is at times reflected by post-depositional vesiculation in surge deposits (e.g. Lorenz 1974).

What, then, are the potential sources of the fluids providing grain-support in surges? As for pyroclastic flows (Section 7.2), the obvious sources are magmatic volatiles (both those initially present and those that diffuse from juvenile fragments within the surge), external water incorporated during phreatomagmatic eruptions, and volatiles derived from vegetation over which surges travel. Although the role of external air ingested into the head and fluidising the flow as a significant aid to grain support for the bulk of the solids has been discounted above, it is likely that significant mixing of air into the billowing top of the surge occurs (Allen 1982). There, the mixing will provide turbulent support for only the finest grainsizes transported by the surge.

## 7.9 Depositional processes in surges

Any transporting medium begins to deposit its load when the energy levels drop below the threshold level needed to maintain discrete grain support or tractional entrainment, or drop below the momentum of a mass of particles moving *en masse* as one. Surges, although initiated by an 'explosive' thrust, once in motion, behave largely as gravity controlled density currents (Allen 1982). As the turbulence which provides grain support declines in response to cooling and slowing, a changing mode from suspended transport, associated initially with an *erosion stage*, to surface traction transport will commence. The energy that generates turbulence for grain support is, as described above, of two types: convective turbulence, due to the heat in the surge, and velocity-induced turbulence in a medium of low viscosity and high lateral velocity. Turbulence due to fluidisation-induced streaming of external air is considered to be minor (Section 7.8).

Convective turbulence will be dissipated as the surge cools in transit. This will occur as latent heat is lost during condensation of steam to water and as cold air is mixed into the surge, particularly along its top. However, convective turbulence will probably be subordinate to velocity-induced turbulence. As the velocity declines, the sediment-carrying capacity of the surge will also decrease, and the surge as a density current will be dissipated. Allen (1982, p. 397) proposes four causes for the dissipation of a sediment density current with a steady head and uniform body:

(a) density reduction through sediment loss or engulfment of ambient medium,
(b) reduction in flow thickness, related to flow stretching or collapse,
(c) friction at the base and upper surface of the flow, partly involving density reduction due to mixing between flow and medium and
(d) reduction in bed slope.

Fall-out of sediment will cause a decrease in the bulk density of the surge and loss of momentum ($= mv$, mass × velocity), and hence velocity and turbulence, according to the Reynolds Number criterion. A reduction in the slope will also lower the velocity, as potential energy is being lost. Given the low density of surges, the coefficient of friction between the surge and its substrate will be low while the surge is inflated and carrying a large suspended sediment population. However, when the surge begins to deflate significant grain fall-out will produce a concentrated basal tractional grain layer. There will be a higher coefficient of friction between this grain layer and the substrate, permitting sedimentation and further slowing of the surge.

Frictional retardation of a surge can also occur by mixing of air at the head and top of the surge. Allen (1982) suggested that the dissipation rate of a density surge is

$$\text{dissipation rate} = h_b/L \qquad (7.11)$$

where $h_b$, the thickness of the body of the surge, is equal to $0.01H$, $H$ being the height of the head, and $L$, the streamwise length of the mixing zone, is equal to about $2$–$3H$. The dissipation rate is in the order of 0.01–0.001 (Allen 1982). Mixing of air into the top probably occurs, but will have little effect on the density of the head and body. If surges are similar to turbidity currents in a *general* fluid dynamic sense, then the body should flow more quickly than the head, implying that once solids have reached the head they should be circulated upwards into the head proper, where mixing with the ambient medium occurs due to the growth of Kelvin–Helmholtz instability waves (billows) on the top of the head (Allen 1982, p. 400). The upper part of the head is diluted by this, and becomes less dense and therefore slower.

For relatively dry surges, grain transport will pass from suspended and surface traction modes of transport to increasing degrees of tractional transport. The *rate* at which grain transfer from the suspended to the traction mode occurs will control whether relatively low grain concentration tractional transport and sedimentation occurs (producing corresponding structures such as cross-stratification and well defined lamination), or whether high grain concentration shear-induced grain-layer flow occurs, with attendant high degrees of particle interaction, deposition by freezing of the grain mass and deposition of a massive bed. It can be speculated that with wet surges the *condensation of steam* should also be an important influence in causing sedimentation. *Adhesion* occurs as fine ash and water droplets begin to adhere to each other, producing clumps. The effective transported grainsize will be increased by this. Condensation should therefore lead to an increased rate of sedimentation compared with the pre-condensation rate of sedimentation, because the greater effective transported grainsize may exceed the prevailing competency of the surge, and because of the adhesion effects between the substrate and moist particles in the surge. The rate at which condensation takes place will therefore be important.

If surges do segregate into a lower concentrated ground layer and a more dilute suspension cloud, then some deposits should contain a separate 'body' layer and an associated fall-out layer (e.g. Fig. 7.35).

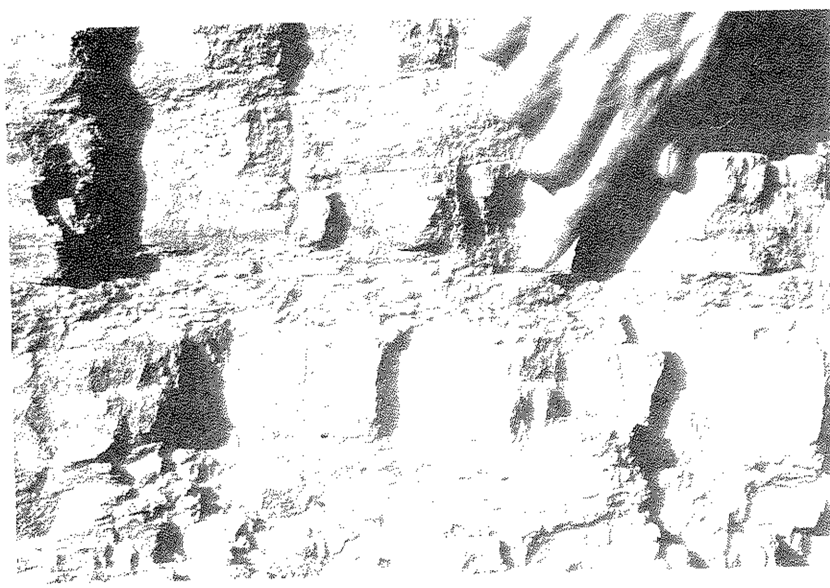

**Figure 7.35** Probable distal rhyolitic base surge deposits from a Devonian tuff ring-caldera lake succession, Snowy River Volcanics, eastern Victoria. The massive layers represent surge 'body' deposits separated by accretionary lapilli-rich 'co-surge' ash fall layers (finger).

## 7.10 Facies characteristics of surge deposits

The facies characteristics of surges can be readily discussed in terms of some of the essential facies parameters introduced in Chapter 1, including:

- geometry
- grainsize
- sorting
- shape and vesicularity
- composition
- depositional structures

### 7.10.1 GEOMETRY

The geometry of surge deposits will depend on the type of surge, the topographic control and post-depositional erosion. *Base surges*, being a vent-related facies, will build up an annulus around the vent which is wedge-shaped in cross section, thinning radially away from the vent. Base surges form one of the principal facies of tuff rings and maar volcanoes (Ch. 13). The deposit from a single surge will be a thin sheet, with minor variations in thickness being controlled by pre-depositional topography and the surface bed-form (e.g. dune forms). However, base surge deposits are almost invariably composed of multiple layers representing multiple surge events (e.g. Figs 5.21 & 22). Such a pile may be tens of metres thick around the vent (e.g. Crowe & Fisher 1973, Schmincke *et al.* 1973, Sheridan & Updike 1975, Fisher 1977). Individual layers may be up to a metre thick near the vent, although usually closer to 50 cm or less, and distally away from the vent, only millimetres to several centimetres thick. Base surges probably do not flow further than 10 km from the vent, and usually only several kilometres. As such, a succession of base surge deposits in the rock record is indicative of proximity to the vent. They may contain intercalated air-fall layers, as well as thin near-vent pyroclastic flow deposits (e.g. Fisher *et al.* 1983). Successive base surge deposits may be separated by thin co-surge ash-fall deposits up to several centimetres thick, derived from a trailing ash cloud (cf. co-ignimbrite ash-fall deposits).

*Ground surges*, usually being expulsions from the head of a pyroclastic flow will usually deposit thinner facies intervals than base surges do (e.g. Fig. 5.23). They may be no more than several metres thick, and usually a metre or less, and should be directly overlain by the basal facies of the parent pyroclastic flow. Being ejected from a moving pyroclastic flow, the geometry and extent of ground-surge deposits will largely be controlled by the topography into which the pyroclastic flow has moved.

*Ash-cloud surge* deposits should occur as a thin (up to several metres thick) sheet of fine ash, mantling the deposit of the host pyroclastic flow behind which it trailed (e.g. Fisher 1979; Fig. 5.21). However, the preservation potential of ash-cloud deposits is low because of the effects of post-depositional erosion. Unless quickly buried beneath further eruptive products, they are likely to be stripped off shortly after emplacement.

### 7.10.2 GRAINSIZE

The grainsize of surge deposits reflects both the degree of fragmentation at the time of eruption *and* the competency of surges to carry particular grainsizes (Ch. 1). The coarseness and variance in the grainsize is thought to be a reflection of the levels of turbulence (G. P. L. Walker 1983). As Crowe and Fisher (1973) pointed out, because surge-forming eruptions are highly pulsatory and of variable explosive strength, there may be considerable changes in the flow-power and in the grainsize characteristics of successive surges, and even within single surges. Near to the vent ballistic fragments and air-fall materials may also be incorporated into surge deposits. However, both *base-surge* and *ground-surge* deposits, because of their high energy state and because their sources at the vent and the heads of pyroclastic flows may contain considerable coarse debris, are capable of carrying, and do carry, significant amounts of large lapilli-size clasts. However, *ash-cloud surges*, being the products of continued elutriation processes in the head and body of pyroclastic flows, will be low energy systems and their deposits will be fine-grained, rarely containing lapilli, and then probably only highly vesicular, low density ones.

## 7.10.3 SORTING

The sorting characteristics of the different surge deposits will also be variable, and will be controlled by the degree of fragmentation, the density variations in the debris being carried, the turbulence levels, adhesion processes and the time available for sorting. As discussed in Chapter 1, the sorting is an hydraulic process, and in populations containing grains with a variety of size, density and shape, variation in grainsize will not necessarily reflect poor hydraulic sorting. Nevertheless, given the origins of the three surge types, it is likely that *ash-cloud surge* deposits will be better sorted than *base-surge* or *ground-surge* deposits. In addition, because of the relatively low particle concentration in surges, the sorting will be significantly better than in most pyroclastic flows, but less so than in fall deposits (Fig. 7.36). However, field observation shows that major grainsize variations occur between the layers in a single surge deposit, especially base-surge deposits, suggesting a highly pulsatory mode of flow and grain transport within a single surge. Secondly, sorting is generally still much poorer than in most normal sedimentary systems transporting equivalent grainsizes, even in turbidites, the nearest sedimentary analogue.

## 7.10.4 SHAPE AND VESICULARITY

The shape and vesicularity of grain populations in the deposits of the three surge types will be more a

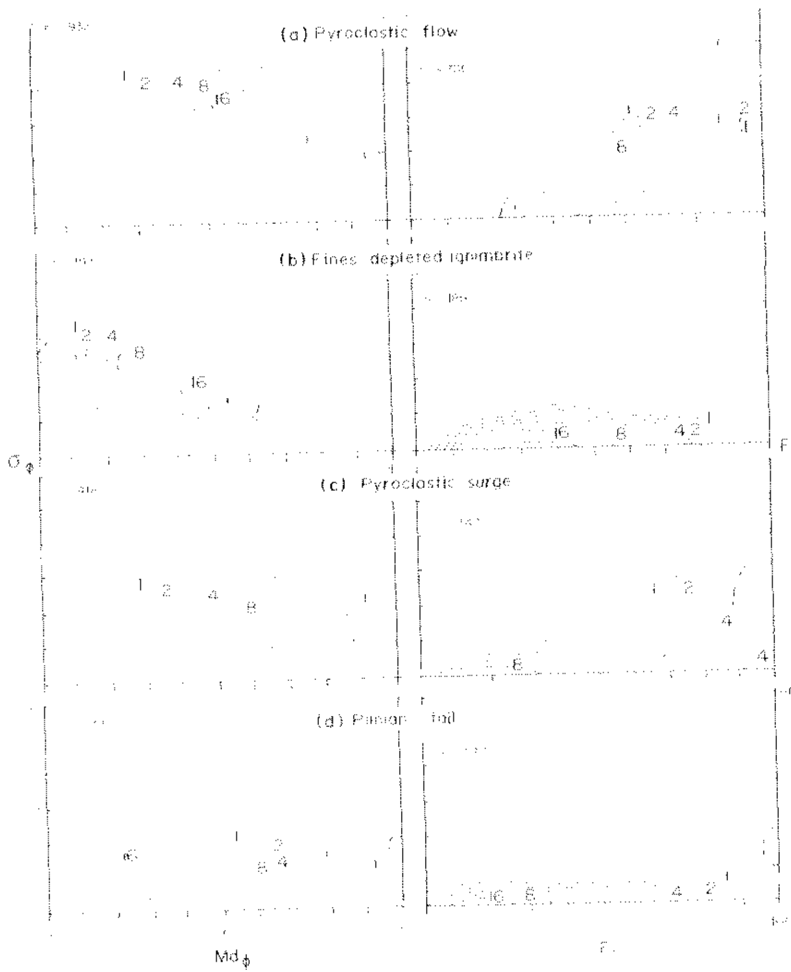


**Figure 7.36** Plots of $Md_\phi$, $\sigma_\phi$ and of weight percentage finer than [illegible] mm ($F_2$) versus weight percentage finer than 1 mm ($F_1$) for pyroclastic flows and three kinds of fines depleted pyroclastic deposits. In (b) various kinds of fines depleted facies associated with ignimbrite are [illegible] ground layers, fines depleted ignimbrite and gas segregation pipes. [illegible] (after [illegible] Walker 1983).

reflection of the mode of fragmentation of the magma during eruption than of flow processes. Hence, base-surge deposits, being of phreatomagmatic origin, will be dominated by poorly vesicular, blocky fragments where the erupting magma is poorly vesiculated (Ch. 3), as will the deposits of ground and ash-cloud surges whose host pyroclastic flows have been associated with phreatomagmatic eruptions (e.g. Self 1983). However, ground and ash-cloud surge deposits derived from eruptions driven by magmatic explosions (Ch. 3) will contain abundant vesicular fragments, although concentration processes during the initiation of ground surges may concentrate denser lithics and crystals in the surges (Section 5.7).

## 7.10.5 COMPOSITION

The composition of the erupting magma and products has an indirect relationship to surge types. Basaltic pyroclastic eruptions essentially do not produce pyroclastic flows. As a result, essentially all basaltic surge deposits found in the rock record can be inferred to be near-vent base surge deposits. The converse is not true, however. Intermediate and silicic eruptions can give rise to all three surge types: base, ground and ash cloud. Accidental clasts may be a significant element in a surge deposit due to explosive incorporation at the vent, and particularly in phreatomagmatic base surge deposits. Accessory clasts may be picked up in transit.

## 7.10.6 DEPOSITIONAL STRUCTURES

Depositional structures are diverse in surge deposits, and have been recognised as a response to varying flow and physical conditions ever since surges were recognised as a pyroclastic transportational and depositional agent (J. G. Moore *et al.* 1966, J. G. Moore 1967, Fisher & Waters 1970, Waters & Fisher 1971, Heiken 1971, Crowe & Fisher 1973, Schmincke *et al.* 1973, Mattson & Alvarez 1973, Sheridan & Updike 1975, Fisher 1977, Wohletz & Sheridan 1979, Allen 1982, Leys 1982, Fisher *et al.* 1983, Edney 1984, Edney & Cas *in prep.*). Many of these authors have also recognised lateral changes in the nature of bedforms or associated internal depositional structures, or both, with distance from the vent, as a response to changing flow conditions as dissipation of surge energy occurs.

As with normal sedimentary structures, structures in surge deposits can be classified as:

*pre-depositional*,
*syn-depositional and*,
*post-depositional*.

### *Pre-depositional structures*

Pre-depositional structures include *erosion gullies* and *U-shaped channels* (Ch. 5) carved out of the depositional surface and infilled by surge deposits. The origin of these depressions may be normal epiclastic erosional processes or erosion by pyroclastic surges (e.g. Fisher 1977, Richards 1959, J. G. Moore 1967; Ch. 5.7, Fig. 5.21c & d). Others include *planar slide surfaces*, usually on the inner crater wall where segments of the unconsolidated, frequently bedded pile of pyroclastics forming the upper crater wall collapse and slide back into the crater (Fig. 7.37). Younger pyroclastic surge and fall deposits may accumulate on this slide surface.

### *Syn-depositional structures*

Syn-depositional structures include wave-like (or dune-like) structures called *dune forms* or dunes, and their internal *cross-stratification*, *massive beds* without structure and *planar beds*. In a discussion of the lateral facies changes of surges from vent to

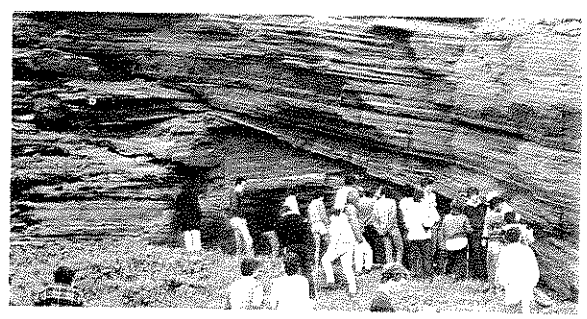

**Figure 7.37** Slide surface sloping into the crater of the Lake Purrumbete maar tuff ring, Western Districts, Victoria, Australia. Younger surge deposits have been deposited on this slide surface. Flow direction is from right to left.

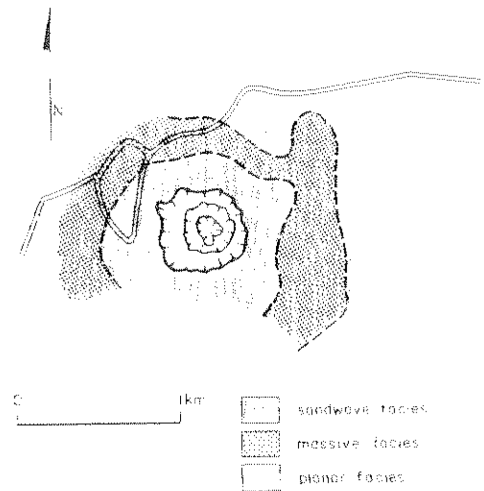


**Figure 7.38** Distribution of major surge facies relative to distance from vent, Ubehebe Crater, California, USA. (After Wohletz & Sheridan 1979.)

distal settings. Wohletz and Sheridan (1979), using many of the concepts presented by Sheridan and Updike (1975), document the occurrence of these syn-depositional structures relative to distance from the vent, based on studies of many volcanic centres in the western USA. Facies intervals dominated by wave- or dune-like bedforms and cross-stratification occur closest to the vent, facies intervals with dominantly massive, structureless beds occur at medial distances and facies intervals dominated by planar beds are most distal (e.g. Fig. 7.38).

Wohletz and Sheridan (1979) suggest that the ordering of these facies with distance from vent can be related to the changing flow conditions within a surge with time and distance, and the implication is that a facies model, much like the Bouma sequence for turbidites, can be considered. The documented grainsize characteristics for these facies (Wohletz 1983, Sheridan & Wohletz 1983) indicate that all are dominated by ash-sized material (Fig. 7.39). However, Edney (1984) and Edney and Cas (*in prep.*) have recognised that there is also an extensive assemblage of lapilli-dominated base-surge facies in the hydrovolcanic basaltic centres of western Victoria, Australia, including massive to diffusely layered lapilli facies (Fig. 7.40), cross-bedded lapilli facies (Fig. 5.21a) and planar-bedded lapilli facies (Fig. 7.40). These are associated with equivalently structured ash facies, presumably similar to those of Wohletz and Sheridan (1979). The massive to diffusely layered lapilli facies at least, is considered by Edney and Cas (*in prep.*) to represent a proximal near-vent surge or surge grain-layer underflow, perhaps deposited by the head of the surge as suggested by Levs (1982), or both. $Md_{\phi}/\sigma\phi$ plots (Fig. 7.41) show that the massive lapilli facies are better sorted than pyroclastic flows (cf. Fig. 7.36) and that all facies are coarser than the previously defined facies of Wohletz and Sheridan (cf. Fig. 7.39). We suggest that, given the complexities of surge mechanics (dry, hot, wet, cold, condensation, intensity of fragmentation, variation in degree of explosive thrust), it may yet be premature to propose simple proximal to distal and upsequence facies changes and models similar to the Bouma facies model for turbidites. Lapilli and ash-surge facies have quite different hydraulic equivalence, and the relationships are bound to be complex.

The terms dune form, dune and antidune have

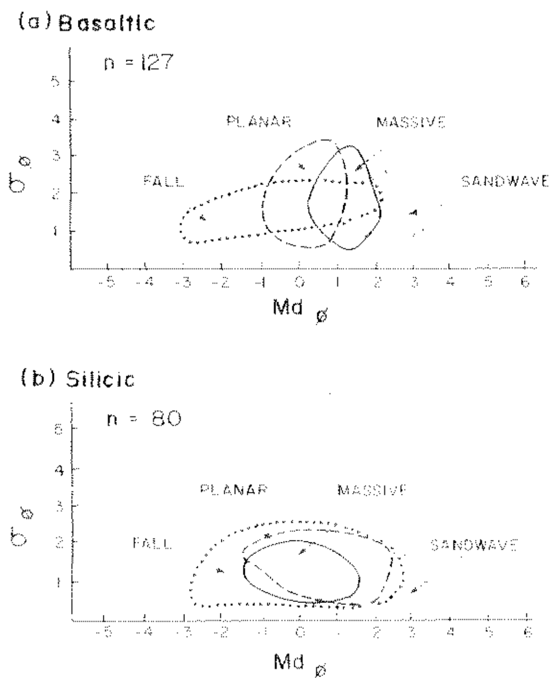


**Figure 7.39** Grainsize characteristics of the principal surge facies of Wohletz and Sheridan (1979) compared with associated fall deposits. (After Wohletz 1983.)

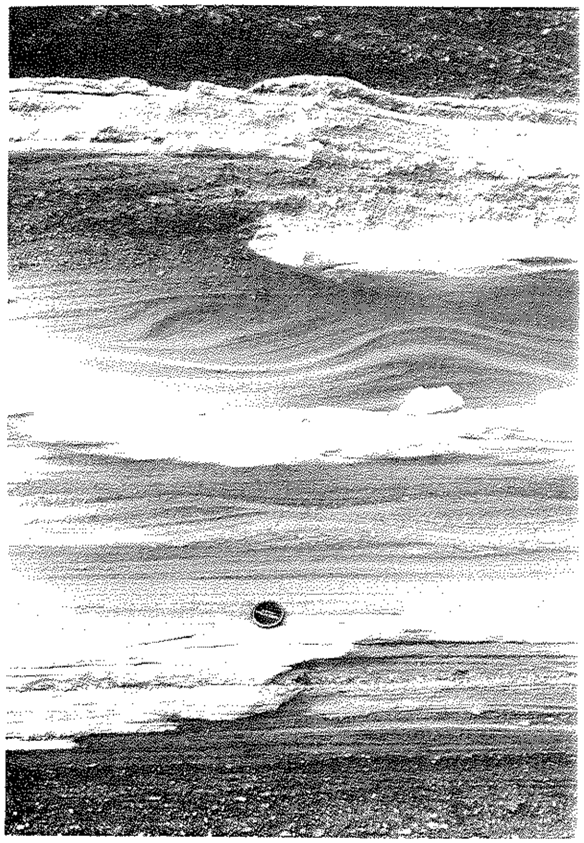

Figure 7.40 Three depositional facies of base surges: dune form (top), planar bedded (middle) and massive to diffusely layered lapilli (bottom). The upward succession from massive to planar bedded to dune-form facies reflects progressively lower concentration surges passing this point. The corresponding decrease in grainsize corresponds with increasing degrees of fragmentation and explosiveness or decreased surge competency, or both. Note the thin lenses of very low angle cross-stratified ash within the planar bedded facies representing incipient dune form formation. The dune-form facies interval (top) contains cross-beds ranging from less than angle of repose, to steeper than angle of repose, reflecting the wet cohesive nature of the phreatomagmatic ash. Flow is obliquely into the photograph from right to left. Holocene Tower Hill Volcanic Centre, Western Districts, Victoria, Australia.

all been applied to the wave-like structures that characterise **dune-form facies** (both ash and lapilli deposits, Figs 5.21, 22 & 7.40). Allen (1982) points out that 'dune' and 'antidune' have been used almost invariably synonymously with *aqueous* epiclastic wave forms. However, the use of these terms

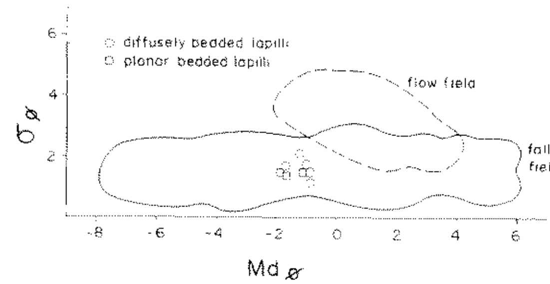


**Figure 7.41** Grainsize characteristics of massive and diffusely layered lapilli surge facies, Tower Hill Volcanic Centre, western Victoria. (After Edney 1984.)

for surge bed-forms may be incorrect, because the physical processes involved in the formation of pyroclastic dune forms and aqueous dunes and antidunes are quite different. Dune-like forms may be symmetrical or asymmetrical. Asymmetry may be either upflow or downflow. Dune-form bedding set thickness decreases approximately logarithmically away from the vent (Wohletz & Sheridan 1979) and average wavelength also decreases away from the vent (J. G. Moore 1967, Waters & Fisher 1971, Wohletz & Sheridan 1979). Wavelength and wave height are logarithmically proportional to each other (Allen 1982).

Cross-stratification angles are highly variable (Figs 5.21, 22 & 7.40). Although they approach the angle of repose in some instances, in most instances they are considerably less than the angle of repose, suggesting that the origin of the cross-stratification is not analogous to low flow regime, aqueous, dune formation, in which cross-beds form by *passive*, inertial gravitational avalanching of grain layers down the downstream face of the dune. Such a process also operates with the formation of aeolian dunes, and some ripples and cross-stratification. A similar process may be responsible for high angle cross-stratification in surge dune forms, probably associated with dry surges.

*Low angle cross-stratification* is due to high velocity, current-driven grain layers being sheared over an irregular, waved surface which itself has a relatively low relief due to high bed shear stress operating. Low angles of cross-stratification in surge deposits have frequently been equated with aqueous antidunes, irrespective of whether the dip direction has been towards the source (normal with aqueous antidunes) or away from the source. The implication of this is that upper flow regime conditions, analogous to those in aqueous unidirectional flow systems, were operating. However, as Allen (1982) points out, this direct analogy is not valid. In the aqueous system we are dealing with cohesionless sediment entrained within an aqueous medium with a density of 1 g cm$^{-3}$. In the pyroclastic surge situation we are dealing with particulate matter driven by a fluid which has a density very much less than that of water.

In the case of *dry surges*, which are simple, *two-phase gas–solid systems*, the processes may be analogous to normal aeolian processes. With dry surge dune forms, normal asymptotic, angle of repose cross-strata of avalanche origin may occur but, even so, the competency of surges in transporting coarse debris is well beyond the limits of normal surface winds, and low angle cross-stratification is common. In the case of *wet, three-phase surge systems* there seems to be no direct analogy with normal sedimentary processes and systems, so implications of analogous flow regime conditions and processes should strictly be avoided. With wet surges, added complications arise because of the existence of a changeable three-phase system, the adhesion processes taking place in transit and the cohesive nature of the substrate over which the surge travels. The last two of these are likely to be responsible for the often very fluidal nature of the layering in surge dune forms. The depositional process will involve adhesion interaction between entrained detritus and substrate, and deposition will take place under the influence of a 'smearing' bed shear force.

Further support for the view that flow conditions in surges and normal sedimentary systems are not analogous comes from the very significant grainsize variations between the layers of the same surge cross-bed set, suggesting a much more pulsatory non-equilibrium flow system. This grainsize variation can be from ash to lapilli sizes. Nevertheless, there is conceptual, and probably some fluid dynamic, equivalence between aqueous antidunes and surge dune forms, given that the low angle cross-stratification requires high shear stresses at the bed surface – much higher than that associated with low flow regime, high angle of repose cross-stratification.

Nevertheless surge dune forms display sufficient regularity in form for them to be used in determining surge transport directions. Observations on modern volcanoes have shown that dune crests are continuous and are orientated perpendicular to the surge flow direction (J. G. Moore 1967, Waters & Fisher 1979, Allen 1982; Fig. 5.19d). Internal cross-stratification can also be used to infer flow directions (Figs 5.21–23 & 7.40). Allen (1982) related the documented variations in bedforms and cross-stratification relationships in base surge deposits to sedimentation rate, surge temperature and moisture content (Fig. 7.42).

In this scheme he has recognised three categories of surge dune bedform: *progressive*, *stationary* and *regressive*. In progressive bed-forms (types $A_1$, $A_2$ and B, Fig. 7.42) the crests of the dune forms migrate in the direction of surge flow. Where the sedimentation rate is high, the internal arrangement of cross-strata resembles normal sedimentary climbing ripples and cross-stratification (type $A_1$). Progressive dune forms are considered by Allen to be the result of surges that are relatively dry or hot, or both. Stationary dune forms (type F) are distinguished by crests which migrate neither upstream or downstream. Regressive dune forms are distinguished by an upstream migration of the crest of the dune form, as represented in the crestal point of successive sigmoidally-shaped cross-strata in a single set. Allen suggests that regressive dune forms are the products of wet or cool surges. Although the crests are retreating upstream, the through-flow of solids within individual layers is downstream. In regressive bed-forms, the depositional process involves cohesive interaction between the transported debris and the substrate. Coarser fragments occur on the upstream side of the crest (cf. progressive dune forms). 'Climbing' effects, in this case upstream, result from both high sedimentation rates and adhesion–plastering due to wetness. Allen suggests that stationary dune forms result from surges with temperatures at the condensation or boiling point of water (Fig. 7.42). In support of his scheme, Allen (1982) points out that

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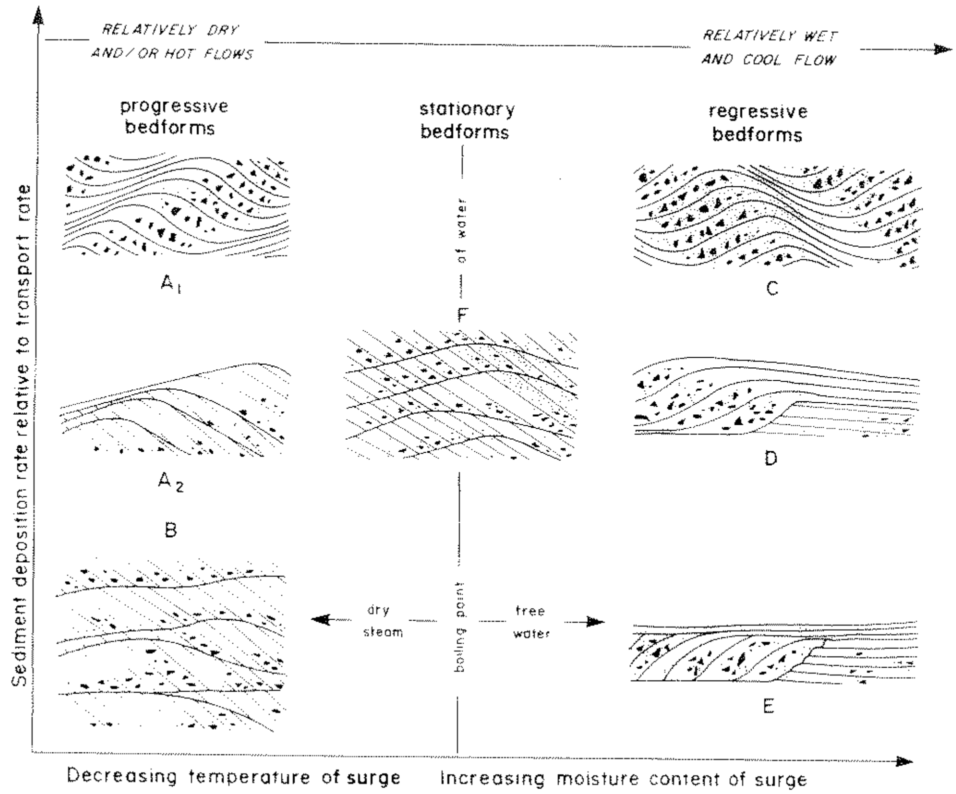


**Figure 7.42** Classification of base surge bedform and internal cross-stratification variations related to depositional rate (relative to transport rate; vertical axis) and surge temperature and moisture content (horizontal axis). (After Allen 1982.)

only deposits with stationary or regressive dune forms are known to contain accretionary or armoured lapilli, or both, as well as vesicles which are due to post-depositional expansion of trapped hot volatiles in wet, impermeable ashes.

Small amplitude, irregular adhesion ripples, similar to those formed in wind-swept snowfields, are another, less common, form of dune forms associated with surge deposits (Allen 1982, Leys 1982).

The cross-stratification in surge dune form facies can be very variable in form. In high angle of repose sets it can be upward concave, asymptotic and wedging downstream. In other situations it is much more fluidal, showing significant thickness variations and, unlike normal sedimentary cross-stratification, individual layers may be continuous from one dune form into the next, showing pinch and swell geometry, and thinning over the crests of waveforms. The dips in cross-beds may vary. Where it is greater than the angle of repose (>30–35°), adhesion processes or soft-state plastic deformation, or both, perhaps under the shearing influence of the host surge, are almost certainly responsible. Angle of repose cross-stratification may also be due to these influences or due to lee-side avalanching from a dry surge or from part of one. More commonly, however, surge cross-stratification is characteristically of a low angle, indicating the influence of significant bed shear or wet state smearing influences, or both, rather than grain avalanching. Although the cross-beds in individual sets may be conformable with each other, very low angle, and irregular, truncations may be common, and these reflect the highly pulsatory nature of surges, even during deposition.

Massive ash facies (cf. massive lapilli facies, above) are considered by Wohletz and Sheridan

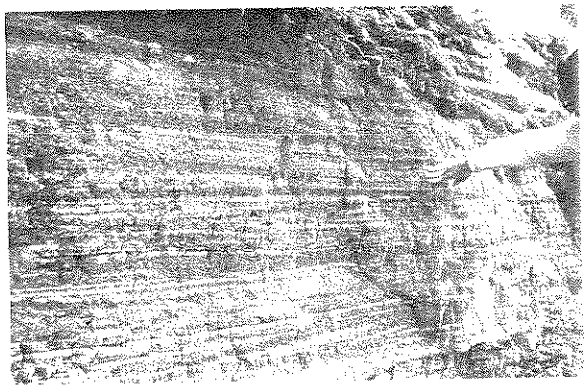

**Figure 7.43** Very low angle cross-stratification interval (middle) within planar bedded facies of uncertain but probable surge origin, Eocene–Oligocene Waiareka–Deborah volcanics, Oamaru, New Zealand.

(1979) to be transitional between dune form and planar bedded facies. They describe beds of massive facies as being frequently lens-shaped and occurring on the lee-side of dune forms. They are ungraded, and generally unstructured internally except for occasional planar to wave-like diffuse layering or pebble trains. Wohletz and Sheridan (1979) consider massive beds to be deposited by a deflating, highly concentrated stage of surge flow, perhaps involving some fluidisation. The massive character is indicative of high particle concentrations and very rapid rates of deposition, probably involving *en masse* freezing of the grain population. The diffuse layering probably reflects internal shearing at the time of deposition in the highly concentrated grain aggregate.

Planar bedded facies consists of layers millimetres to several centimetres thick which are planar, slightly wavy or even locally lensoidal. Just as in the cross-stratified sets of dune forms, major grainsize variations are common from layer to layer and between different sets. Some are ashes, others lapilli deposits (Edney 1984, Edney & Cas *in prep.*). Layers are generally laterally extensive, and reverse grading is common. The grainsize variation between layers again suggests a highly pulsatory depositional process during the formation of a set of planar beds. Reverse grading indicates that a layer of grains is shearing laterally over the substrate, larger grains migrating upwards out of the zone of maximum shear at the interface between the surge and the substrate. Grain fall-out from suspension is not indicated by this reverse grading. The reverse grading, being of a shear origin, also implies that the surge is in a deflated, highly concentrated state at the time that this, the most distal facies recognised, forms. Planar-bedded facies may be difficult to distinguish from bedded fall deposits without evidence of lateral transport (scours, cross-stratification, low angle truncations).

**Figure 7.44** Cross-stratified surge deposits viewed perpendicular to flow direction showing wavy layering, and low angle cross-stratification and truncation, Mt Leura, western Victoria, Australia.

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4 Cross-stratified
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Planar beds may be *gradational* with cross-stratified intervals, and in some instances may have a *very* low angle inclination (Fig. 7.43) appearing to be planar but being incipient cross-beds (W. Edney *pers. comm.*). However, in this regard the perspective is important. Exposures that are perpendicular to flow direction will expose wavy, parallel to subparallel layering of cross-bed sets (Fig. 7.44).

*Post-depositional structures*

Post-depositional structures commonly associated with surge deposits include *bomb sags* (Fig. 5.21g & h), formed when ballistic blocks and bombs lob into wet, unconsolidated surge deposits producing *soft-sediment plastic deformation*. *Flame structures* at the base of surges have also been described (Crowe & Fisher 1973) and *soft-sediment oversteepening* of dune form cross-strata under the shearing influence of the succeeding surge is common. Such oversteepening may bring strata that are normally at lower than angle of repose inclinations (<30°), to angle of repose (30–35°) or even steeper. The layering is usually still intact, but is stretched and squeezed due to plastic deformation while in a wet, cohesive state. Similarly, normal soft-sediment slumping of wet, cohesive ash is common (Fig. 5.21f).

## 7.11 Surges compared with turbidity currents

In Section 7.6 it was suggested that, as a crude generalisation, surges are akin to turbidity currents in the normal sedimentary sphere, whereas pyroclastic flows are akin to viscous debris flows. This general analogy is based on the fact that surges and turbidity currents are turbulent, gravity-controlled mass flows with low particle concentration, whereas pyroclastic flows and debris flows are generally both viscous, laminar-plug flow systems. There the analogy ends, however.

Whereas turbidity currents begin to flow because of their potential energy, surges are initiated by an explosive thrust, with the exception of ash-cloud surges. Thereafter there is a complex energy chain, as discussed in Section 7.7. Whereas in turbidity currents the particulate population is essentially cohesionless, this is clearly not the case with surges (Sections 7.8–10). Turbidity currents are simple two-phase solid–liquid systems in which the density contrast between the solids and the supporting liquid is relatively small compared with that in the complex three-phase (solid–liquid–gas) system of most surges. The relative proportions of the three phases may change in surges, as may the effective dynamic granulometry, due to adhesion processes leading to clumping.

Because of all these differences, hydrodynamic equivalence between turbidity currents and surges cannot be considered valid (Allen 1982). However, once initiated, both move as gravity-controlled, turbulent, cloudy masses, which seem to produce a systematic succession of facies during deceleration and deposition. The facies are not the same, and in both cases not all facies elements need be produced by all flows. Much work still needs to be done on the facies and facies relationships produced by surges.

## 7.12 Pyroclastic surges and pyroclastic flows – relationships

The discussion of pyroclastic flows and surges so far has emphasised their unique characteristics, the former being described as poorly inflated, non-turbulent, concentrated mass flows which maintain their kinetic energy over long distances and long periods (relatively), whereas the latter highly inflated, turbulent, low particle concentration, short-term phenomenon which rapidly dissipates. However, it has also been shown (Section 7.5) that there are some extremely violent pyroclastic flows which are at least partly turbulent, and are capable of producing wavy stratification, bedforms and low angle truncations and cross-stratification (Fig. 7.30) in their veneer deposits. The Taupo ignimbrite (G. P. L. Walker *et al.* 1981b, C. J. N. Wilson 1985), the Rabaul ignimbrite (G. P. L. Walker *et al.* 1981c) and the ignimbrite of the 1815 eruption of Tambora (Self *et al.* 1984) are key examples. This brings into question the degree to which pyroclastic flows and surges are distinct entities

(Section 5.6). Each is a type of flow, but is there a spectrum from dense pyroclastic flows (e.g. nuée ardentes, block and ash flows) at one end to dilute surges at the other? G. P. L. Walker (1983) and G. P. L. Walker and McBroome (1983) touched on this problem. In both these publications Walker poses the possibility that two of the better-documented so-called surge events of modern time – those of Mt Pelée in 1902 and the 18 May 1980 cataclysmic eruption of Mt St Helens, were not surges, as has been asserted in the literature to date (Fisher *et al.* 1980, A. L. Smith & Roobol 1982: Mt Pelée; Hoblitt *et al.* 1981, J. G. Moore & Sisson 1981: Mt St Helens; Section 5.6), but highly violent pyroclastic flows. Waitt (1981) observed that the characteristics of the 18 May Mt St Helens deposit were similar to neither purely surge nor pyroclastic flow deposits, but shared the characteristics of each. Hoblitt *et al.* (1981) and J. G. Moore and Sisson (1981) also noted anomalies compared with the documented characteristics of surge deposits. The diverse descriptions and subdivisions of the so-called surge deposits by different authors make it difficult to extract a consensus stratigraphy but, following Walker and McBroome, it seems that three principal layers can be identified.

Layer 1 is a relatively well sorted gravelly or sandy layer with rapid lateral changes in thickness and grainsize and some lateral discontinuity. Grainsize and thickness decrease outward from the source, and significant fines depletion is evident. Layer 2 contains two facies, a massive one which 'has the aspect of a pyroclastic flow deposit particularly where it occurs in valley ponds' (G. P. L. Walker & McBroome 1983, p. 571), and one showing stratification and dune forms. The massive facies is homogeneous, occurs in valley pond settings, and contains gas escape pipes where it is thick. Layer 3 is very fine, contains accretionary lapilli, is rarely more than several centimetres thick and is agreed by all authors to be an ash fall-out of dilute trailing ash clouds derived from the 'blast-

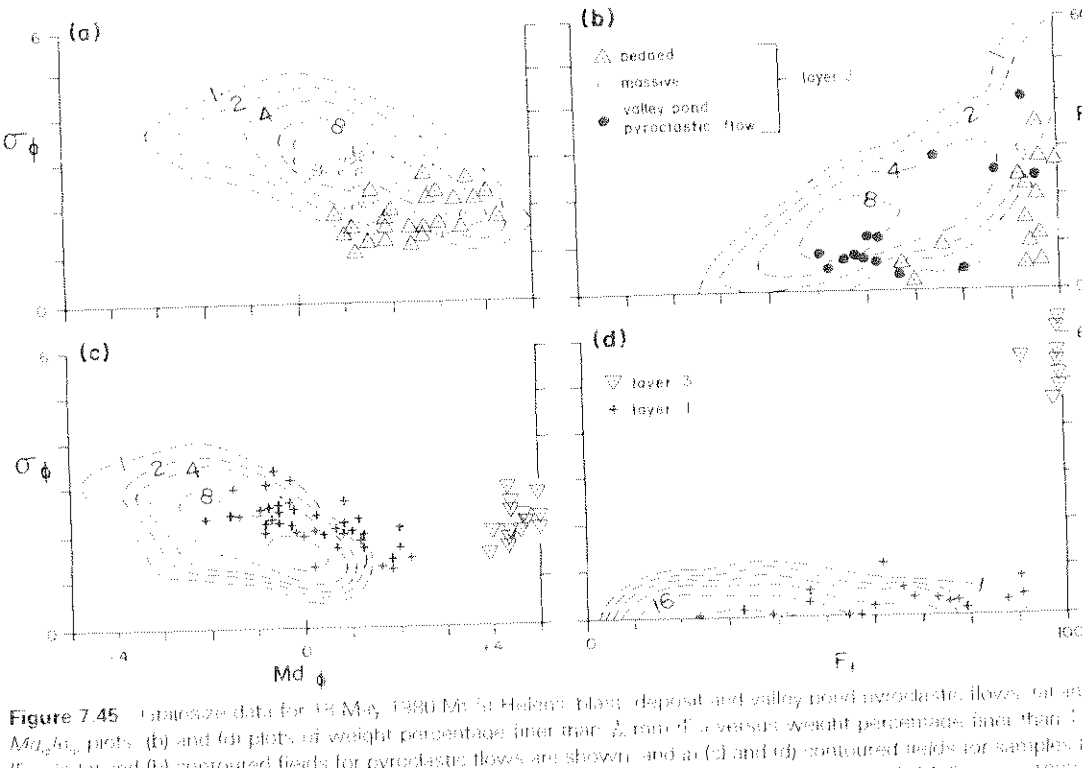


**Figure 7.45** Grainsize data for 18 May 1980 Mt St Helens 'blast' deposit and valley pond pyroclastic flows. (a) and (c) $Md_\phi/\sigma_\phi$ plots; (b) and (d) plots of weight percentage finer than 1/16 mm ($F_2$) versus weight percentage finer than 1 mm ($F_1$). In (a) and (b) contoured fields for pyroclastic flows are shown, and in (c) and (d) contoured fields for samples from fines-depleted facies of ignimbrite (G. P. L. Walker 1983) are shown. (After G. P. L. Walker & McBroome 1983.)

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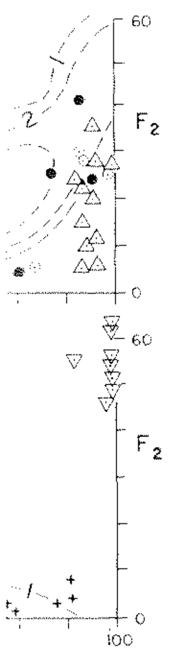



flows (a) and (c)
finer than 1 mm
for samples from
roome 1983.)

surge' body. The relationship between layers 1 and 2 is generally abrupt, but locally their interface is gradational (Moore & Sisson 1981, Waitt 1981). The total thickness of this assemblage of layers rarely exceeds 1 m, except in hollows.

Based on $\sigma_\phi$ versus $Md_\phi/\sigma\phi$ plots and plots of $F_2$ (the weight percentage finer than 1/16 mm) versus $F_1$ (the weight percentage finer than 1 mm) plots, G. P. L. Walker and McBroome (1983) show a close relationship between the layer 1 and 2 samples and pyroclastic flow fields (Fig. 7.45). However, samples show a significant depletion in fines, indicating better sorting than most pyroclastic flow deposits.

Similarly, G. P. L. Walker and McBroome (1983) point out that the surge deposits of Fisher *et al.* (1980) and Fisher and Heiken (1982) associated with the 1902 Mt Pelée eruption consist of three layers: a basal gravelly to sandy layer, a stratified to massive middle layer and a fine, thin, accretionary-lapilli bearing, ash layer. Grainsize characteristics are similar to those of Mt St Helens.

G. P. L. Walker and McBroome (1983) and G. P. L. Walker (1983) interpret both the Mt St Helens and Mt Pelée deposits to be the products of very violent, low aspect ratio pyroclastic flows. In the Mt St Helens case this is based on the grainsize affinities with pyroclastic flows, and the doubt that dilute surges could travel 30 km and maintain turbulent suspension of clasts over that distance. Layer 1 is interpreted as the fines-depleted ground layer deposited from the fluidised head. The dune forms and their internal wavy layering seem ill-defined relative to most true surge deposits, suggesting weak, low levels of turbulence (G. P. L. Walker 1983). The main layer, layer 2, seems very similar to valley pond ignimbrite, and also occurs on gently sloping valley walls, as a thin veneer deposit (Ch. 8; G. P. L. Walker & McBroome 1983).

It is possible then that two modern catastrophic events which had been interpreted as surges (Section 5.6), may have been very violent, even partly turbulent pyroclastic flows, similar to the historic ones erupted from the Taupo, Rabaul and Tambora volcanic centres. It is therefore implicit that a spectrum of pyroclastic flow types exists, from very dense, high particle concentration ones to relatively low concentration, but violent, ones. There is an implication in this that there is also a spectrum of grain-support processes from perhaps inertial grain flow avalanching through fluidised, laminar semi-plug flow (Sparks 1976, C. J. N. Wilson 1980), through to fluidised, turbulent flow (C. J. N. Wilson 1980). Given that surges are turbulent, but that during deflation they may undergo shear-induced laminar flow, there is not much of a quantum gap between dilute types of pyroclastic flow and surges, implying a nearly complete spectrum of processes and deposits. However, on that point G. P. L. Walker (1983) maintains that discreteness is required because, to maintain integrity and to be capable of transporting large dispersed clasts, pyroclastic flows are dependent on fluidisation, which can only be maintained while *fines are retained* in order to act as resistance to, and to channel, fluidising ingested air. In surges, fines loss is inherent because of the high level of turbulence and the consequent removal of large volumes of fines by elutriation, which essentially discounts the possibility of fluidisation. On the other hand, however, fines loss or depletion can also occur in certain situations related to pyroclastic flow processes (and not just surge transport; G. P. L. Walker 1983; Fig. 7.46) without affecting the overall coherence of the pyroclastic flow. It is clear that much is yet to be learned about the processes operating in pyroclastic flows and surges and how these two flow processes are related. This is more than vindicated by the diverging views on the G. P. L. Walker and McBroome (1983) interpretation, as shown by the responses and reply to that paper (Hoblitt & Miller 1984, Waitt 1984, G. P. L. Walker & Morgan 1984)!

## 7.13 Further reading

The papers by Sparks (1976), Sheridan (1979), Fisher (1979), C. J. N. Wilson (1980, 1984) and C. J. N. Wilson and Walker (1982) are essential reading on aspects of flow and depositional processes in pyroclastic flows, even though they have been summarised here. On surge processes and facies

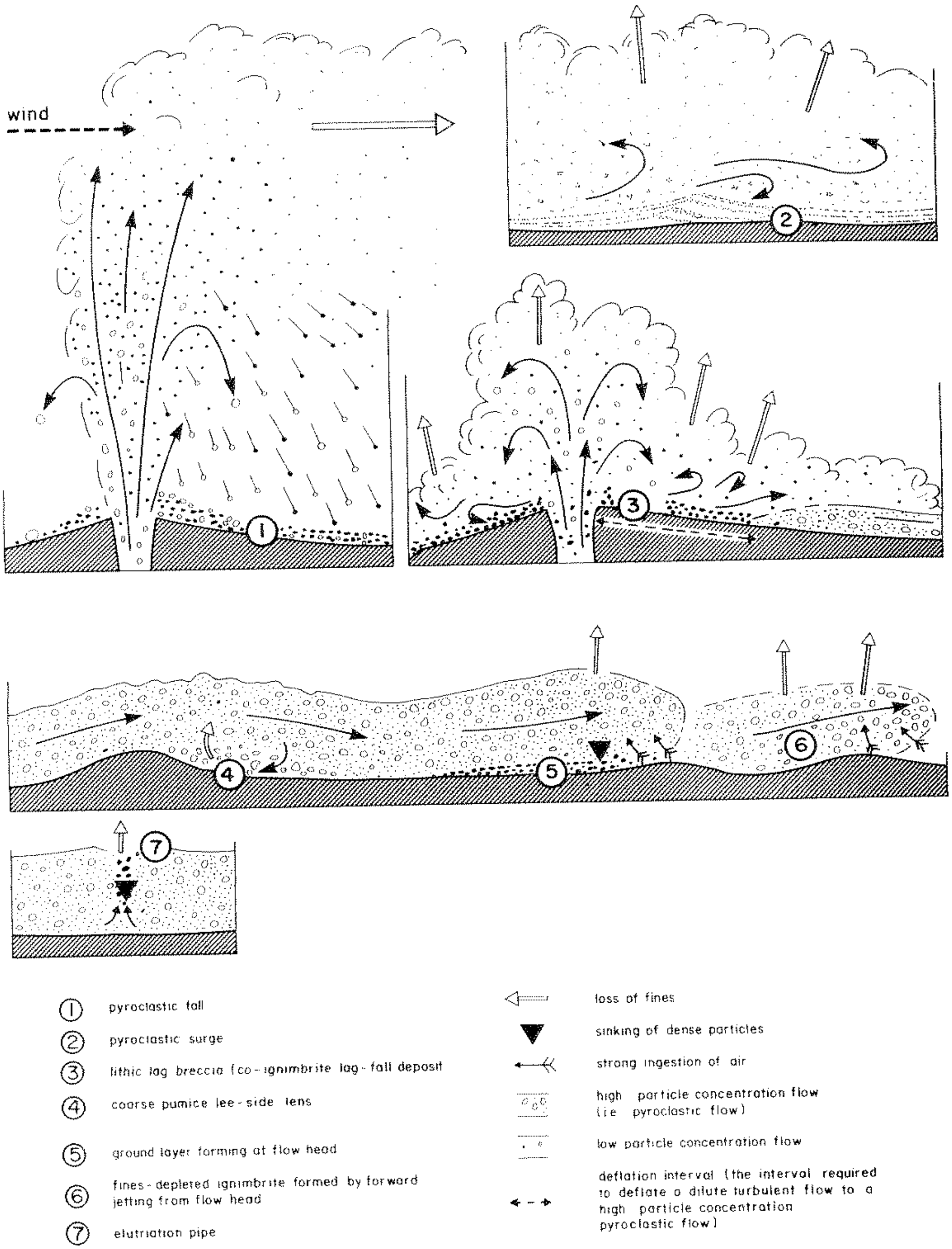
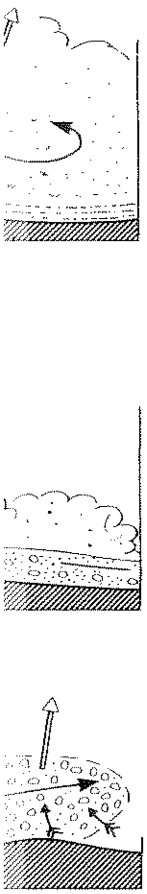



**Figure 7.46** Schematic views showing seven situations which produce fines depletion and good sorting of pyroclastic deposits. (After G. P. L. Walker 1983.)



characteristics, the review by Wohletz and Sheridan (1979) is extremely useful, as are the earliest papers on surge deposits by J. G. Moore (1967), Fisher and Waters (1970), Crowe and Fisher (1973) and Sheridan and Updike (1975). On the problem of the distinction between violent pyroclastic flows and turbulent surges, the papers by G. P. L. Walker (1983) and G. P. L. Walker and McBroome (1983) are recommended.