Basaltic-volcano systems

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Abstract: This review and personal account of basaltic volcanism concentrates on the geological aspects and physical controls, unlike most others which concentrate on geochemistry. It serves as an update to reviews by Wentworth & Macdonald 20 to 40 years ago. The concept of volcanic systems is developed, a system including the plumbing, intrusions, and other accoutrements of volcanism, as well as the volcanic edifice. Five types of systems are described, namely lava-shield volcanoes, stratovolcanoes, flood-basalt fields, monogenetic-volcano fields, and central volcanoes (having silicic volcanics as well as basaltic). It is postulated that the system-type depends on (a) the magma-input rate and (b) the frequency with which magma-batches enter the system (or modulation frequency). These controls determine whether a hot pathway is maintained from the source to the high-level magma chamber, and hence if eruptions are concentrated in the central-vent system. Reviews are given on many aspects: the different kinds of rift zones that volcanoes exhibit, and their controls; the mostly small intrusions, including coherent-intrusion complexes, that occur in and under the volcanic edifices; the great sill swarms that are an alternative to flood basalts; the origin of the craters and calderas of basaltic volcanoes; high-level magma chambers, their locations, and cumulate prisms; the positive Bouguer gravity anomalies that are attributed to cumulates and intrusion complexes; the structures of basaltic lava flows; the characteristics of basaltic explosive volcanism and the consequences of participation by non-volcanic water; the products of underwater volcanism; the origin and significance of joints, formed either by contraction or expansion, in volcanic rocks; and the distribution of vesicles in basaltic rocks. A fairly extensive bibliography is included.

This is in part a review; it takes stock of changes in concepts since the publications by Wentworth & Macdonald (1953) and Macdonald (1967), which were models of clear description of basaltic volcanics, and since the publication 12 years ago of the landmark Basaltic Volcanism Study Project (1981). It also incorporates new ideas, new interpretations, and new ways of looking at the subject. The scope of basaltic volcanism is too broad to cover all aspects, and petrology and geochemistry are specifically omitted.

More than half of the world's volcanoes are basaltic or include basalt among their products; and about one third of known eruptions, involving about 20 volcanoes per year erupt basaltic magma. Basaltic volcanoes occur in all tectonic settings. Basaltic volcanism is associated with both divergent and convergent plate boundaries; it characterizes spreading ridges mostly concealed beneath the ocean; it characterizes hotspot volcanism which, in oceanic settings, is almost exclusively basaltic and, in continental settings, is commonly bimodal (basaltic plus rhyolitic); and it is widespread associated with andesitic and more silicic magmas in subduction-zone settings, particularly where oceanic lithosphere is subducted below oceanic or thin continental lithosphere, as in island arcs (e.g. the Mariana and Kurile arcs).

Basaltic magma is derived by incongruent partial melting of mantle peridotite, favoured in tectonic settings (e.g. hotspots and rifts) where mantle rock rises adiabatically to relatively shallow levels, or in subduction-zone settings where volatiles decrease the melting temperature of mantle rock.

Magma properties are basic parameters in volcanology:

- magma density relative to lithosphere density makes volcanism possible and helps determine the positions of magma chambers and intrusions;
- viscosity and yield strength determine the geometry and structures of lava flows and intrusions;
- gas content promotes eruptions and determines their explosivity; and
- gas content combined with viscosity and rheology controls the explosive violence of eruptions by determining the ease with which gases escape from magmas.

Little mention is made of these basic magma properties, because their values tend to be rather uniform amongst different basaltic volcanoes. They are, thus, the unifying features of basaltic volcanism. More attention is therefore directed at the influence of non-magmatic variables (e.g.
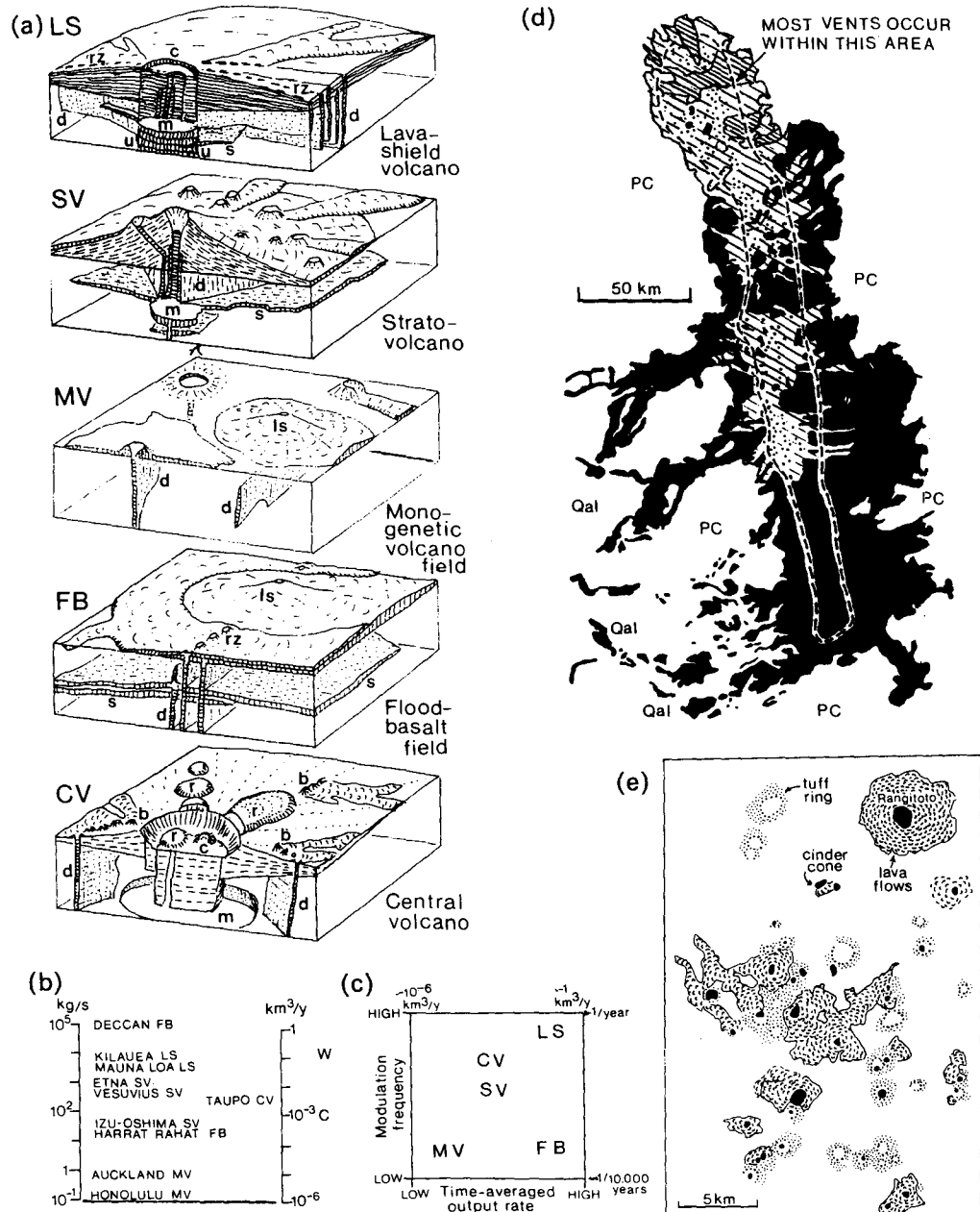
Fig. 1. Volcano types and volcano systems.

(a) The five types of basaltic-volcano systems, illustrated schematically by block diagrams. b: basaltic vents; c: caldera; d: dyke; ls: lava shield of scutulum type; m: magma chamber; rz: rift zone; r: rhyolitic lava dome; s: sill or intrusive sheet; u: cumulates. (b) Time-averaged output rates of selected volcanic systems. W: equivalent world energy consumption by man (excluding food); C: equivalent energy consumption (including food) by city having a population of 1 million. (c) Inferred fields for the five volcanic-system types on a plot of modulation frequency against time-averaged output rate. (d) Sketch map of a moderate-sized flood-basalt field: Harrat Rahat in Saudi Arabia (Camp & Roobol 1989). Solid black: lavas of 10–1.7 Ma old; open diagonal shading with dots: lavas of 1.7–0.6 Ma old and their cinder cones; close diagonal shading: lavas of <0.6 Ma old, including those of AD 641 and 1256; PC: Precambrian basement rocks; Qal: Quaternary sediments near Red Sea coast. (e) Sketch map of the Auckland monogenetic-volcano field, New Zealand, simplified by omitting the coastline (after Searle 1964).

magma-supply rate and involvement of non-magmatic water) in causing most of the diversity.

Volcano types and volcanic systems

Five volcano types are distinguished (Fig. 1a): lava-shield volcanoes, stratovolcanoes, flood basalts, monogenetic volcanoes, and central volcanoes, this last having a significant proportion of silicic products in addition to basalt. Flood basalt vents, as well as monogenetic volcanoes, erupt once, and once only. Lava shields, stratovolcanoes and central volcanoes are polygenetic structures that erupt more than once.

There is much merit in regarding volcanoes as parts of magmatic or volcanic systems. A system may embrace the intrusions, magma chambers, conduits, magma source and accompanying geothermal fields as well as the volcano itself. The concept of volcanic systems acknowledges that the visible edifice of a volcano is only one part of a bigger entity.

A volcanic system may be compared with a system of civilization such as a city, with the extensive infra-structure of transportation, water and power supply, waste disposal and communications on which a city depends. Just as a volcanic system is sustained by a supply of energy (in the form of hot magma), so a city system is sustained by a supply of energy (in the form of fossil fuels, electric power, and food). Very roughly, a city of 1 million people requires about the same power input as a volcanic system having a magma input of 1 million m^3 per year.

Basaltic systems have a source in the mantle from which magma ascends, mainly because of its positive buoyancy but sometimes aided by tectonic forces, toward the surface. They have one or more conduits by which the magma ascends. Polygenetic volcano systems generally possess a high-level magma chamber, situated at a neutral buoyancy level, which stores magma and modulates its delivery to the volcano and to sub-volcanic intrusions. Deep storage reservoirs may also exist.

Lava-shield volcanoes

These consist mainly of lava flows and have the form of low-angle shields. Slope angles tend to be small, mostly 4°–15° although steeper examples are known. Rift zones tend to be narrow and well-defined but grade into radial vent systems. The eruptive fissures are marked by spatter ramparts and mostly-small cinder cones.

Lava-shield volcanoes, as discussed here, are polygenetic structures commonly exceeding 1000 km^3 in volume, and are not to be confused with small monogenetic lava shields categorized by Noe-Nygaard (1968) as of scutulum type (Latin: *scutulus*, diminutive of *scutus* – 'shield') which have volumes of 0.1–15 km^3.

Examples of lava-shield volcanoes Mauna Loa in Hawaii is the largest. It has an estimated volume of 40 000 km^3, rises to 9–10 km above the surrounding ocean floor and has a keel that projects well below it. Subaerial slopes are mostly 3–6° except in the scars of old landslides. Pico in the Azores and Fogo in the Cape Verdes have steep (> 30°) cones atop the shield. Shields in the Galapagos consist of an upwardly-convex dome with slopes as steep as 30° rising above the low-angle shield, and some of the eruptive fissures that occur on a narrow plateau around their large summit calderas are annular. Some lava shields, notably Kilauea and perhaps also Pico, consist mainly of pahoehoe. Others, such as Mauna Loa, consist of roughly equal proportions of pahoehoe and aa. Others again, for example Madeira, Tutuila (Samoa) and the Galapagos volcanoes are composed mainly of aa.

Stratovolcanoes

These consist of a stratified succession of lava flows and interbedded pyroclastic deposits and tend to have a conical form with a slope that generally steepens upward until approximating the repose angle (about 33–36°) of loose debris. Most subduction-related basaltic volcanoes conform with this type. Commonly the cone is truncated by a caldera. Rift zones tend to be diffuse and ill-defined and grade into radial vent systems, and cinder cones tend to be conspicuous features on the volcano flanks. In many arc volcanoes basaltic andesite and more silicic types accompany basalt.

Examples of stratovolcanoes The Japanese island volcano of Izu-Oshima is representative of the arc-type basaltic cones. It has a basal diameter of 27 km, rises about 2200 m above the ocean floor to 758 m above sea-level, and has a volume of 415 km^3 (Suga & Fujioka 1990) of which 23 km^3 are above sea-level. It has subaerial slopes of up to about 20° but the cone is truncated by a caldera about 3 km in diameter. Pyroclastic rocks are more voluminous than lava flows. A broad and rather ill-defined rift zone parallels the long-axis of the island. It is marked by cinder cones and, near the coast, phreatomagmatic tuff-rings.

Izu-Oshima has displayed periodic activity through the past 10 000 years with on average about 100 years between larger eruptions (Tazawa 1984). From the careful volumetric study by Nakamura (1964) of eruptive products in the historic period, Izu-Oshima is a prime example of a 'steady-state' volcano, in which magma is fed at a uniform rate into the magma chamber and the output is modulated by the chamber characteristics.

Fuji, the highest volcano in Japan, rises to 3776 m above sea-level and about 3700 m above its base and has an estimated volume of 1400 km^3. It is a fine example of a stratovolcano having the form of a typical andesite cone, but is exclusively basaltic apart from a minor volume of dacite pumice erupted explosively in the latest (1707) eruption (Tsuya 1955). The basal diameter is 25 km. The steep part of the cone above the 2000 m level has slopes exceeding 20°, but a broad apron sloping mostly <10° below the 1500 m level accounts for three-quarters of the total area.

Monogenetic volcanoes

These consist of clusters of scattered and mostly small (>2 km^3) volcanoes, each generated by a single eruption. Most commonly a volcano consists of a cinder cone associated with outflows of aa lava, but some are lava shields of scutulum-type (e.g. Rangitoto Island, Auckland, and Xitle in Mexico), and many that occur near the coast or close to lakes are phreatomagmatic tuff-rings or maars.

Some monogenetic fields (as Auckland) are exclusively basaltic whereas in others some of the volcanoes are more silicic. Thus the Kaikohe field north of Auckland includes a rhyolitic lava-dome. The Higasha-Izu field is bimodal; about 50 out of >70 volcanoes are basaltic, and the others are andesite, dacite or rhyolite (Aramaki & Hamuro 1977; Hayakawa & Koyama 1982). In some fields only a small proportion is basaltic. Xitle for example is the only basaltic volcano in the Chichinautzin field just south of Mexico City (Martin del Pozzo 1982).

Examples of monogenetic-volcano fields Two examples of young monogenetic fields are described from contrasted tectonic settings, namely Auckland, situated behind an active arc system on continental crust, and Honolulu in a hotspot setting on oceanic crust. Both are small-scale examples. In both fields, future eruptions may be expected, at long time intervals. The individual volcanoes may be considered as dead, but the volcanic systems are alive. ·

The Auckland field (Fig. 1e) is one of several in the northern part of New Zealand (Searle 1964; Heming & Barnet 1986). It coincides rather closely with the extent of Auckland City and consists of an apparently random scatter of about 50 basaltic vents spanning roughly the past 60 000 years. The setting is one of a drowned landscape in young eroded sedimentary rocks. Surface water or groundwater participated in many eruptions. More than half of the volcanoes are maars and the rest are cinder cones and lava flows. The latest (770 years BP) and largest eruption built Rangitoto Island, a low-angle shield of scutulum type, 6 km in diameter and 260 m high at the central cinder cone. The volume above sea-level is 1.2 km^3.

The Honolulu Volcanics monogenetic field on Oahu, Hawaii, includes tuff-rings of Diamond Head and Hanauma Bay, and the Salt Lake and Aliamanu craters which contain lherzolite and garnet pyroxenite mantle-derived xenoliths. Some volcanoes are clearly distributed along fissures up to 4 km long.

The eroded vent systems including volcanic plugs of ancient polygenetic fields are known in many places, for example, in Fifeshire and the Edinburgh area in Scotland (Geikie 1897), and the Hopi Buttes in Arizona.

Flood-basalt fields

These consist of monogenetic volcanoes erupted from widely scattered vents, but their lava flows cover wider areas than in monogenetic-volcano fields, overlap or are superposed to form parallel-stratified successions, and have much greater volumes.

Giant flood-basalt fields have volumes in the range 10^5–10^7 km^3 (Yoder 1988; White 1992). They are distributed through geological time at average intervals of 32 Ma (Rampino & Stothers 1988), and each one formed at the time of inception of a hotspot, on arrival of an ascending mantle plume at the asthenosphere/lithosphere boundary (Richards *et al.* 1989).

Examples of flood-basalt fields Examples are the 16 Ma Columbia River Basalts in the northwestern USA (Tolan *et al.* 1989), and the 69–65 Ma Deccan Traps of peninsular India that may be implicated in the biological mass-extinction event at the Cretaceous/Tertiary boundary.

Moderate-sized flood-basalt fields related to the Ethiopian hotspot occur east of the Red Sea/Dead Sea rift system from Yemen to Syria. They include Harrat Rahat in Saudi Arabia (Camp & Roobol 1989; Fig. 1d). It spans the past 10 Ma and the youngest flow was erupted in AD 1256

(Camp *et al.* 1987). Future eruptions may be expected, at long time intervals. The field has a wide scatter of cinder cones marking the hundreds of vents, and the locus of activity tended to migrate northward with time. Several flows that travelled down valleys toward the Red Sea are about 100 km long. Lavas cover 20 000 km^2 and have an estimated total volume of 2000 km^3.

Other flood-basalt fields occur along the eastern part of Australia. In one, the McBride Province in Queensland (Stephenson *et al.* 1980), lava flows cover 6000 km^2 and came from a wide scatter of vents. The flows span 3 Ma and the youngest is the 190 ka Undara flow that is 170 km long and has a volume between 10 and 23 km^3. This flow is pahoehoe and has an exceptionally long system of lava tubes along its length (Atkinson *et al.* 1975; Atkinson 1991).

Many flood-basalt fields overlapping in time and space occur in Iceland. The basalts erupted from rift zones typically 10–20 km wide that can be traced 40 to >100 km across country. Some systems have a central volcano on the rift zone and may be better characterized as central-volcano systems. The several en echelon rift zones of Reykjanes lack central volcanoes. Each rift zone is about 40 km long by 7–15 km wide and has eruptive fissures in the central part and non-eruptive fissures on the periphery.

Central volcanoes

These are stratovolcanoes or shield volcanoes that have a significant proportion of silicic volcanic rocks in addition to basalt (usage of Johnson 1989), and generally have a bimodal composition in which rhyolite and basalt predominates and rocks intermediate in composition are scarce or absent. The silicic rocks form pumice and ash-fall deposits, ignimbrites, and stubby lava flows or lava domes. Very commonly the volcanoes have one or more calderas resulting from subsidence consequent on large silicic eruptions.

Examples of central volcanoes Newberry volcano situated 60 km east of the Cascade Range in Oregon is a broad shield-like edifice with flanks inclined mostly under 4° capped by a steeper cone. The flanks are diversified by about 400 cinder cones and fissure vents arranged in several rift zones feeding extensive lava flows. Compositions include basalt although basaltic andesite predominates. The cone is truncated by a caldera 7 × 5 km in size, shallow because it is partially infilled by silicic flows, domes and pyroclastic deposits. Compositions include rhyolite, but rhyodacite predominates. Outflow sheets of ignimbrite occur on the flanks. Newberry has a shadow zone 12 km wide within which basaltic vents are absent but outside which they are abundant. The volcano has an estimated volume of 450 km^3 and erupted six times in the Holocene (Higgins 1973; Chitwood 1990).

Jebel Kariz in southern Arabia is a fine example of a central volcano the internal structure of which is revealed by deep erosion (Gass & Mallick 1968).

Volcano morphology

The steeply conical form of many stratovolcanoes is due to several causes, notably the prevalence of low-intensity eruptions that tend to pile the erupted material close to the vent, and the magma viscosity which tends to be higher on stratovolcanoes than on shield volcanoes and increases the explosivity (Fig. 2).

The form of lava-shield volcanoes reflects the high proportion of output released on flank rift zones instead of at the summit, and the generally high discharge rate during eruptions which tends to produce lava flows that travel far. Caldera collapse and a general subsidence concentrated in the summit area also contribute to the morphology.

Volcano collapse

Destructive processes have an important effect on volcano shape. Basaltic volcanoes can build to very large structures several kilometres high, and if they are built on sediments their foundations are weak. The dip of the layers outward from the centre and the presence of pyroclastic layers and hydrothermally altered zones produce inherent weakness in the volcanic edifice. Dyke injections that forcibly shoulder aside the rock to make space for themselves, local updoming of central volcanoes by the ascent of silicic-magma diapirs, and severe marine erosion on exposed coasts of island volcanoes all conspire to produce instability and cause major failure of parts of the volcanic edifice.

Most large basaltic volcanoes consequently suffer occasional volcano-collapse events. The cone of Stromboli, rising 3 km from the Mediterranean floor, is defaced on the NW side by the collapse scar of the Sciara del Fuoco. Tenerife is scalloped by several collapse scars; that of the Orotavo Valley has a volume exceeding 60 km^3. Las Canadas is possibly a landslide scar and not a caldera. The greatest volcanic landslides are those of Hawaii (Duffield *et al.* 1982; Moore *et al.* 1989). Some were catastrophic events; others not. Some involved 1000 km^3 of rock.
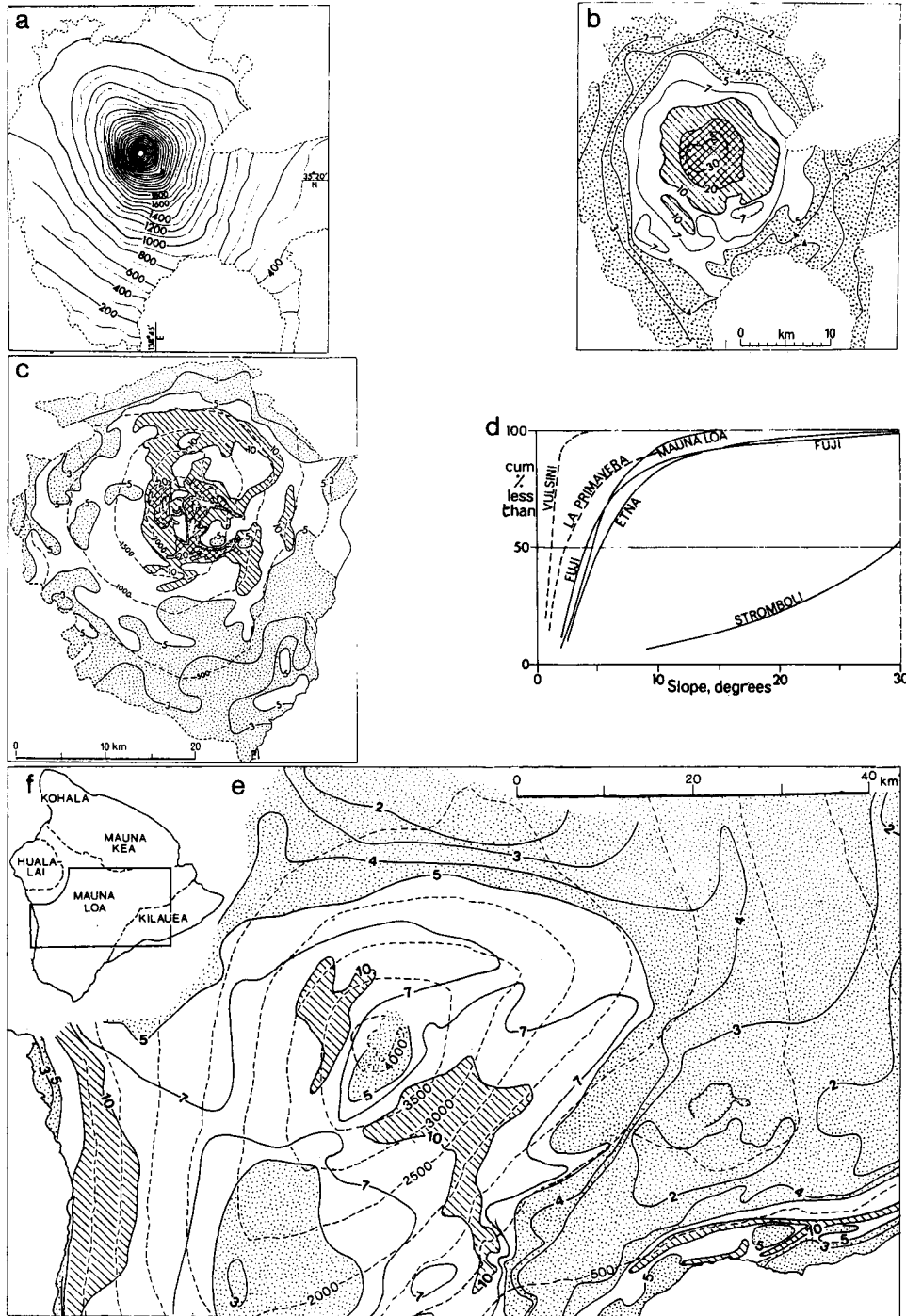
Controls on system type

In the polygenetic volcano systems, magma batches ascend sufficiently frequently along the same conduit that the conduit walls are maintained in a hot condition and provide magma with a thermally and mechanically very favourable pathway toward the surface.

In the monogenetic and flood basalt systems magma batches ascend at such long time intervals that the pathway taken by one batch has effectively cooled by the time that the next batch is ready to ascend. In the absence of a thermally or mechanically favourable pathway, the new batch has to create a new pathway to the surface.

It is postulated here that two principal controls operate to determine the type of volcanic system that develops, namely the time-averaged magma-supply rate and the modulation frequency of the supply.

Consider the supply rate. Strictly it should include the magma that makes intrusions as well as that which erupts, but the former is generally not known and the volcanic output rate is then the best available measure of supply rate. In different volcanic systems the output rate varies over 5 orders of magnitude from $c.\ 1\ \mathrm{kg\,s^{-1}}$ to $c.\ 10^5\ \mathrm{kg\,s^{-1}}$ (Fig. 1b). This rate may alternatively be expressed as an energy flux that varies from below 1 MW to nearly 10^5 MW.

The highest output rates are achieved in the giant flood-basalt outpourings such as those of the Deccan and Columbia River Basalts. The most productive lava-shield volcanoes (as in Hawaii) at the peak of their activity, and some of the more productive flood-basalt fields in Iceland are about an order of magnitude lower. Most stratovolcanoes are one or two orders of magnitude lower still, although some such as Etna are highly productive. The lowest output rates are given by polygenetic-volcano fields.

Fedotov (1981) considered the question of the threshold magma-supply rate below which a hot pathway cannot be sustained and found it to be between 200 and 1600 kg s^{-1} depending on whether the supply is continuous or intermittent, and whether the lithosphere section has a continental or oceanic geothermal gradient. This

threshold agrees well with observed supply rates for monogenetic and polygenetic systems.

Consider now the modulation of the output. Strictly it occurs at a deep crustal or sub-crustal level, and for monogenetic-volcano and flood-basalt systems it is the frequency (commonly 1 per 10^3 to 10^5 years) with which magma batches rise to the surface. For polygenetic volcanoes little firm information exists, and the best that can be done is to take the frequency of eruptions, or of magma excursions from the shallow chamber (Fig. 1c).

The mechanism of modulation at a deep level can only be surmised. Magma that accumulates in or near the asthenosphere source-region exerts an upwardly-directed force because of its positive buoyancy. The magnitude of the force depends on the density contrast between magma and adjacent lithosphere, and the volume of magma that has segregated within a small area. When the force becomes sufficiently great the magma-batch ascends. Monogenetic volcanoes commonly have a volume of 0.1–1 km^3, and give an indication of the volume of magma that is required to create a pathway through the lithosphere. The availability of suitable fractures may also play an important part in aiding magma ascent.

Storage of flood-basalt magma and underplating

A feature of flood-basalt systems is that individual erupted volumes are large so giving a moderate to high time-averaged output rate even though the eruption frequency is low. Because of the low frequency, conduits cool between eruptions and hence flood basalts erupt from a wide scatter of vents, and the individual flood-basalt lavas are monogenetic.

A problem of flood-basalt systems is how and where a large volume of magma is stored before an eruption. The volume that erupts can greatly exceed what is evidently needed to create a pathway to the surface in monogenetic fields. One possible mechanism is that magma accumulates at a deep level of neutral buoyancy (at or near the Moho?) where changes (crystal frac-

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Fig. 2. Volcano types and volcano systems.

(a) Contour map of Fuji volcano. Contour interval 100 m. (b) Slope-angle map of Fuji volcano based on spacing of generalized contours; slope in degrees from the horizontal. (c) Slope-angle map of Etna on the same scale as (a). Dashed lines are contours at 500 m intervals. (d) Curves showing cumulative percentage of map area having less than the given slope angle, for selected volcanoes; La Primavera is a central volcano predominantly composed of silicic volcanics; Vulsini is an ignimbrite-shield volcano; Etna, Fuji and Stromboli are stratovolcanoes. (e) Slope-angle map of part of Mauna Loa on the same scale as (a) and (c); dashed lines are contours at 500 m intervals. (f) Index map of Island of Hawaii. (a)–(c) reproduced by permission of the Royal Society from Volcanological Research on Mount Etna 1975.

tionation; crustal assimilation; gas exsolution) take place that reduce the density of the magma until it becomes buoyant and rises. Alternatively the source region is laterally extensive and lateral connections are good so that, once buoyant ascent begins, magma drains laterally from a wide area into the newly created pathway.

The general absence of mantle xenoliths in flood basalts suggests that the magma resides sufficiently long at some level above the mantle for xenoliths to drop out (Clague 1987). Some continental flood basalts have assimilated crustal rocks, which would be favoured by residence in contact with the crust (Huppert & Sparks 1985). There is also evidence for underplating of the crust, which may be another consequence of deep storage.

Underplating is the formation of mafic intrusions at a deep crustal or sub-crustal level (Cox 1980; Fyfe 1992). Igneous underplating is evidenced by the presence in the lower crust of material having high seismic-wave velocities. Seismic refraction profiles point to the existence of thick underplating, for example, in the North Atlantic region under the sequence of submerged basalts recorded as seaward-dipping reflectors in the upper crust (White 1992). The basalts range in thickness up to 4 km; the intrusions are several times more voluminous.

Underplating may explain why continental crust that is uplifted by a hotspot (e.g. at 140 Ma in Brazil, near the Parana flood basalts; Cox 1989) remains uplifted long after plate motion has carried that crust far from the current hotspot position. Permanent surface uplift occurs provided that the underplating gabbroic rocks are less dense than underlying mantle.

Underplating could occur at a deep level of neutral buoyancy, or alternatively at a level where lithospheric layers having contrasted elastic or rheologic properties are juxtaposed (for example, at a deep crustal level where more rigid rock that tends to behave elastically overlies more ductile rock that tends to flow under stress).

Magnetotelluric surveys in Iceland indicate that a shallow (10 km) partial-melt zone over 100 km wide underlies and extends laterally beyond the active rift zone. This reservoir could store a large volume of magma. During the Krafla intrusive and eruptive events in 1978–1984, chronologically-matching changes took place at the Bardarbunga volcanic centre > 100 km away (Tryggvason 1989) suggesting that some form of lateral interconnection exists.

Unmodulated long-sustained eruptions

Under conditions of high thermal-energy supply rate, a more or less continuous magma pathway may become established between the mantle-source and the surface permitting a long-sustained and unmodulated eruption. In Hawaii such unmodulated activity gives rise to long-lived lava lakes, or generates scutulum-type pahoehoe shields exemplified by that of Mauna Ulu on Kilauea's east rift zone constructed 1969–1974 at a sustained delivery averaging about 5 m^3 s^{-1}.

Compound pahoehoe shields of scutulum-type thought to have a similar origin are common in some flood-basalt fields such as the Deccan, and in the Snake River Plain give rise to what has been referred to as Plains-type volcanism (Greeley 1982; cf. Rutten 1964). Rangitoto Island and Xitle are other examples.

Many scutulum-type shields of relatively primitive basalt developed in Iceland in a short time period at about the end of the Ice Age when the widespread ice sheets rapidly declined. This was a period of greatly enhanced volcanic output, and could be attributed to the melting of the ice effectively reducing the lithosphere thickness by about 5–10% leading to an increase in the degree of partial melting in the upper mantle. Alternatively the redistribution of mass resulting from melting of the glaciers and rise of sea-level flexed the lithosphere and favoured creation of pathways to the surface (Sigvaldason et al. 1992).

Fissure eruptions

In older classifications of basaltic volcanoes great importance was placed on fissure eruptions as a distinctive type. It is now clear that most basaltic eruptions take place initially as a 'curtain of fire' from a fissure, whatever the volcano or system type, but, with time, eruption tends to become concentrated at one or several points. With concentration, local fissure-widening by wall-erosion may occur, enhancing concentration of discharge at that point. Effusion from a point tends to conceal the evidence that initially eruption was along a fissure.

Changes in volcanic systems with time

The active life of most volcanic systems is between 0.1 and 10 Ma and if the magma-supply rate changes significantly during this time or if silicic magma becomes available the system may change from one type to another. These changes provide insights into significant features of the sub-surface plumbing systems of volcanoes.

Well-documented system changes occur in

Hawaii when a decline in magma-supply rate occurs as plate motion conveys a system away from the hotspot focus. Some changes are embodied in the concept that Hawaiian volcanoes pass through a characteristic sequence of life stages (Stearns 1946; details updated by Macdonald *et al.* 1983, Peterson & Moore 1987 and Walker 1990).

The peak of activity of an Hawaiian volcano is the shield-building stage, exemplified by Kilauea and Mauna Loa, when tholeiitic magma is voluminously produced. The declining 'postshield' or 'alkalic cap' stage, exemplified by Mauna Kea, marks a reduction in magma supply rate by one or two orders of magnitude. Magmas are transitional or alkalic basalt. Eruptions tend to be more explosive and build large cinder cones instead of the mostly small spatter deposits of the shield-building stage. Lavas are mostly aa.

Mauna Kea is a shield volcano capped by a stratovolcano, and the wide scatter of cinder cones suggests it may now have become a monogenetic field. The infilling of the summit caldera, the very low modulation frequency of about 1/10 000 years and the abundant inclusions of mafic and ultramafic cumulates indicate that the high-level magma chamber no longer exists, and the well-organized magma-distribution system that channelled magma into rift zones no longer functions.

Fractionation of alkali-cap magma to types such as mugearite and benmoreite undoubtedly takes place perhaps in a deep magma reservoir at the Moho or the base of the volcanic edifice.

Some Hawaiian volcanoes subsequently enter a rejuvenation ('post-erosional') stage in which a monogenetic-volcano field may develop, exemplified by the Honolulu Volcanics on Oahu about 1 Ma following cessation of shield-building activity on Koolau Volcano. Rejuvenation-stage lavas include highly alkalic types such as basanite, nephelinite and melilite-bearing types. Very similar rejuvenation-stage volcanism is seen for example on Lanzarote (Canary Islands) where vents and lava fields of the 1730–1736 and 1824 eruptions are scattered among the eroded stumps of an older volcano (Carracedo *et al.* 1993). Mantle xenoliths are common on Koolau and Lanzarote, indicating that no magma reservoir exists above the mantle-source.

A change of system-type from shield to central volcano is shown by Faial volcano in the Azores. Faial appears to be wholly made up of basalt apart from two trachytic lava domes and a covering up to about 20 m thick of trachytic pumice around the summit caldera. The rocks of Faial are up to 2 Ma old, and the latest eruption was in 1957–8. It is not known precisely when silicic magma first appeared but it likely happened within the past 20 ka.

Volcanism in the Yellowstone hotspot trace exhibits the reverse sequence. The Yellowstone volcano has very infrequent large-volume rhyolitic eruptions (Christiansen 1984) and over the past 16 Ma the locus of rhyolitic activity migrated 700 km northeast into Yellowstone by plate motion (Pierce & Morgan 1992). Some mix-magma lava flows occur, showing that basaltic magma participated in the rhyolitic volcanism. Flood-basalt volcanism of the Snake River Plain, however, followed and lagged behind the rhyolitic activity.

Rift zones, their orientation and intensity

Most basaltic eruptions occur from fissures, and virtually all basaltic volcano systems have eruptive fissures. Fissures are opened very easily by the hydraulic jacking action of magma, and are the 'natural' underground conveyance for low-viscosity magma (Emerman & Marrett 1990). They commonly extend for tens of kilometres and are typically concentrated into rift zones. Magma solidified in fissures forms dykes. Dykes have a high survival potential, and in deeply eroded areas may be virtually all that survives of the volcanic system.

Radial and fascicular fissure distribution

Basaltic volcanoes that possess a magma chamber experience magma excursions which generate dykes or other intrusions and sometimes lead to volcanic eruption. A magma excursion occurs when the wall or roof of a magma chamber ruptures and some magma escapes. Rupture results when, because of magma input or vesiculation, a chamber swells. Magma excursions are critically important in volcano growth, contribute strongly to determining volcano morphology, and power active geothermal systems.

For a volcano that is not buttressed by a neighbouring volcano and occurs in a setting where there is no regional extensional deviatoric stress, the normal consequence of magma excursions is to generate radial fissures.

Radial fissures result from the pressure exercised by a swelling body of magma on its walls, in which extensional deviatoric stresses are set up tangential to the magma-chamber walls. Failure of wall rocks therefore occurs on fractures trending normal to the walls. In the ideal radial swarm the fissures would be straight, and their traces would radiate from a common point. The dyke intensity would be uniform in all directions

at any given distance outward, but because of dyke divergence and restricted travel distance of narrow dykes the intensity would decrease radially outward.

Probably few basaltic volcanoes conform with this ideal; fissures tend to be more concentrated in some sectors than in others, and some fissures curve when traced outward to become approximately parallel. 'Fascicular' (Latin *fascia* – bundle or sheaf, as of sticks or wheat) is proposed for the case where the fissures are concentrated in two sectors 180° apart and tend toward parallelism when followed outward.

The classic example of a radial dyke swarm is that of eroded Spanish Peaks volcano (Colorado; Ode 1957), and Tristan da Cunha is an example of an active volcano having a radial swarm (Chevallier & Verwoerd 1987). Examples of fascicular fissure swarms occur in the Galapagos (Chadwick & Howard 1991), and subduction-related volcanoes that show them include Fuji (Nakamura 1977) Hakone (Kuno 1964), and Shitara (Takada 1988).

Regional stress control

Most polygenetic volcanoes and flood-basalt fields possess rift zones, these being narrow extensional zones in which ground cracks, faults and eruptive fissures are concentrated. Rift zones are underlain by dyke swarms. They vary greatly in their orientation and the spacing of fissures in them.

The orientations of some rift zones are determined by a regional stress field. A good example is Iceland where the rifts are parallel with the spreading axis of the Mid-Atlantic Ridge.

In subduction-related volcanoes that have rift zones, the rifts are orientated normal to the subduction zone and parallel with the plate-convergence direction. Nakamura (1977) and Nakamura *et al.* (1977) regarded them as sensitive markers of regional stress trajectories. Lo Giudice *et al.* (1982) showed that two of Etna's several rift zones conform in orientation with conjugate shear fractures inclined at about 35° on either side of the plate-convergence direction.

The rift zones in that part of the Azores southwest of the Mid-Atlantic spreading ridge are arranged, unlike in Iceland, approximately normal to the spreading axis, and are parallel with horsts and grabens in a major zone of faulting (the East Azores fracture zone) that extends toward the Straits of Gibraltar. This extensional zone, described as a leaky transform fault by Krause & Watkins (1970), is seismically very active and volcanic eruptions tend to be temporally closely related to major earthquakes.

The latest eruption, that of 1957 at Capelinhos off the western tip of Faial, was accompanied by earthquakes and up to 1 m of subsidence along a graben crossing the island (Machado *et al.* 1962).

On Sao Miguel, basaltic eruptive fissures of Holocene age trending northwest as elsewhere in the Azores have an en echelon distribution and occur in a broad and nearly west-trending band of country parallel with the length of the island. These fissures were postulated by Booth *et al.* (1978) to be shear fractures developed by right-lateral transcurrent movement on an underlying west-trending fault.

Gravitational stress control

In Hawaii, eruptive fissures and dykes are tightly concentrated in narrow rift zones some of which are traceable for 50 km on land and up to 70 km more under water. Fiske & Jackson (1973) postulated that they are shallow features, found little evidence for a regional stress or structural control, and attributed these rift zones to gravitational stresses acting on high volcanoes and causing rifting along the elongation-axis of the edifice.

On Etna the edifice effect explains well the orientation of some fissures close to, but outside and parallel with, the edge of the deep Valle del Bove landslide scar (Guest *et al.* 1984; McGuire & Pullen 1989). The marked increase in the rate of fissure formation in recent years possibly has dangerous implications (McGuire *et al.* 1990). Sao Jorge in the Azores also illustrates well the edifice effect. This island is a narrow horst and the precipitous bounding fault scarp along the north is up to 800 m high. Eruptions tend to occur from fissures localized near the crest of the horst and trending parallel with its length.

Note that the extensional stress regime resulting from the edifice effect is likely to be pronounced only at shallow, superficial levels in volcanoes; other controls such as operation of a regional stress field are likely to operate at deeper levels.

Neutral buoyancy control

Impressed by the great number and non-Gaussian intensity distribution of sub-parallel dykes in the Koolau dyke complex, Oahu, Walker (1986, 1992*a*) proposed a mechanism by which concentrations of weakly- or non-vesicular dykes constitute a zone of high density situated in a zone of lower-density vesicular lava flows. Positions of neutral buoyancy exist on both sides and over the top of the high-density zone, and

intrusions tend to be channelled into these positions so accentuating the density zonation. Some earlier dykes moreover have planes of weakness at or parallel with their margins that are commonly exploited by later dykes, so providing an additional reason for localization in a narrow zone.

'Coherent-intrusion complex' was proposed for high-intensity complexes of small intrusions (Walker, 1992a), and two kinds were recognized – namely dyke complexes and intrusive-sheet complexes. Roughly half of documented major basaltic volcanoes possess the latter. Formation of a dyke complex requires that a volcano be capable of widening sufficiently to accommodate the dykes; otherwise an intrusive-sheet (cone-sheet) complex that is accommodated by thickening of the volcano develops instead. Recent studies (Gautneb & Gudmundsson 1992; Walker, this volume) show that cone-sheets are closely similar to dykes in width and in having their dilation vector normal to the intrusion-plane.

Other controls

Walker (1990) observed that along the Hawaiian Islands, rift zones tend to be aligned alternately parallel with the direction of motion of the Pacific Plate over the Hawaiian hotspot, and parallel with major transform faults such as those of the Molokai fracture zone. The faults were planes of weakness exploited by the hotspot magma. Rift zones in Samoa are aligned WNW parallel with the plate-motion direction except on Tutuila where they are aligned ENE, roughly parallel with the North Fiji transform fault. It is thought that an extension on this fault was activated when, by plate motion, the hotspot focus came into alignment with it (Walker & Eyre, in prep.).

Rift propagation rate

In the 1984 eruption of Mauna Loa, rifting began near the summit and fissures quickly propagated some 15 km down the northeast rift zone at an average of 1.2 km h^{-1} (Lockwood et al. 1987). The 1959–60 summit eruption of Kilauea was followed 24 days later by activity 47 km downrift (Macdonald 1962). In its latest eruption (in 1983) a flank fissure on the strato-volcano of Miyakejima extended 4.5 km laterally at 1.5–2.4 km h^{-1} (Aramaki et al. 1986).

From time to time earthquake swarms occur in rift zones in which foci and epicentres migrate progressively down-rift and are thought to mark the lateral propagation of bladed dykes. Epi-

centres in one such swarm in Iceland in July 1978, extended to 27 km from Krafla volcano migrating on average at 1.6 km h^{-1} (Einarsson & Brandsdottir 1980). In another in Kilauea's southwest rift zone in August 1981 (one of many in the 1980s), foci migrated 12 km downrift at 2.6 km h^{-1} (Klein et al. 1987). No eruption occurred in either.

Such seismic swarms are clear, but indirect, evidence for lateral magma excursions. The basaltic eruption of 1874–75 in Sveinagja, Iceland some 50–70 km north of Askja and the simultaneous caldera subsidence in Askja (Sigurdsson & Sparks 1978) is more tangible evidence that magma migrated laterally.

Small intrusions

Volcanic systems characteristically include small intrusions, often in such large numbers as to rival the volume of the volcano (Fig. 3). Intrusions having a narrow sheet-like form predominate and propagation of a sheet-like intrusion by hydraulic fracturing at the advancing tip occurs very readily.

Small intrusions are commonly called hypabyssal, implying a shallow level of emplacement, or subvolcanic, implying that they form beneath where volcanic activity occurs. They are also commonly called minor intrusions, bearing in mind that although individually small, collectively they may be major components of a volcanic system.

The nomenclature of sheet-like intrusions is slightly confused, because it is based on two different criteria, namely dip and whether the intrusions are concordant or discordant to the countryrock bedding. Discordant intrusions that are vertical, or nearly so, are called dykes. Concordant intrusions that are horizontal, or nearly so, are called sills. Most sills, however, locally transgress the countryrock layers at a small angle, typically in step-like fashion. Intrusions that are inclined at a moderate angle either to the horizontal or to the countryrock layers may be called intrusive sheets or inclined sheets.

Individual intrusions are often highly irregular in form. This results from their strong propensity to follow pre-existing planes of weakness such as joints, faults and bedding planes. Side-steps are common where, having followed one joint or plane of weakness, an intrusion steps sideways to follow another.

An important aspect of small intrusions is the attitude of their dilation vector, namely the direction in which the rocks on either side moved apart to accommodate the intrusion. Generally the vector is normal to the intrusion plane or, if
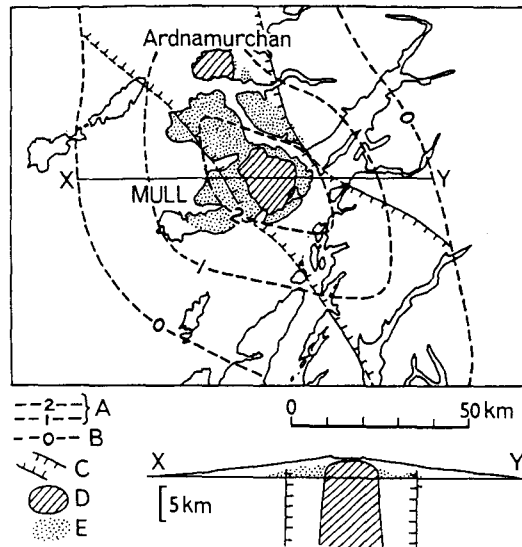
Estimation of original volumes of lavas and intrusions in the Mull Volcano system

(a) Above base of volcano
    Total volume inside B
        9000 km² × 0.9 km (average thickness)   = 8100 km³
    Central intrusive complex (D)
        300 km² × 2 km × 50% of this volume   =   300 km³ ⎫ intrusions
    Dyke swarm outside limits of D   ⎬ 500 km³
        4000 km² × 1 km × 5% of this volume   =   200 km³ ⎭
    Total lavas = 8100 − 500   = 7600 km³
(b) Extra, to an arbitrary depth of 10 km below base of volcano:
    Intrusives (D), mainly cumulate prism
    inferred from gravity anomaly   = 3500 km³
    Dyke swarm (C) outside D
        4000 km² × 10 km × 5% of this volume   = 2000 km³
    Therefore total extra intrusions   = 5500 km³
(c) Total (a) plus (b):
    Lavas   7600 km³ (56% of total)
    Intrusions   6000 km³ (44% of total)
    Total   13600 km³

Fig. 3. Map and section of the Tertiary central-volcano system of Mull, Scotland, and a volume estimate to illustrate the importance of intrusions in such a system. A: 1 and 2 km isopachs for the volcano; reconstruction based on amygdale-mineral zonation; B: inferred original extent of the Mull Volcano – extrapolated line of zero thickness; C: approximate limits of intense dyke swarm; D: central intrusive complexes; E: present extent of lavas of the Mull Volcano.

the intrusion is irregular, to the plane that approximates the average strike and dip of the intrusion (Walker 1987). The dilation vector plunges at a small angle in dykes and at a steep angle in sills.

Some intrusive sheets occur in swarms, the members of which dip toward a common focus, and have been widely termed cone-sheets. A particular mode of origin, namely injection along inverted-conical shear fractures propagated above a magma chamber by upwardly directed magma pressure, tends to be implicit in the definition and the original term 'centrally-inclined sheets' which lacks a genetic connotation is to be preferred.

Most small intrusions are inferred to record magma excursions from a magma chamber that was sufficiently swollen (by incursions of fresh

magma or gas expansion) to suffer chamber-wall rupture. Some excursions lead to volcanic eruptions and the intrusion then marks the pathway to the surface. On Kilauea, however, fewer than half of magma excursions lead to eruption and many other volcanoes experience seismic swarms, with foci ascending with time toward the surface (suggestive of magma ascent in dykes) but without eruption; examples of such 'abortive eruptions' include the Long Valley seismic event of 1983 (Savage & Cockerman 1984).

Bruce & Huppert (1990) showed that narrow dykes can travel only a short distance before they are blocked by rapid cooling. The number and total width of dykes injected in a given time-interval therefore decreases strongly with distance outward from a volcanic centre (Walker 1988). Dykes injected within a given time-interval therefore constitute a wedge tapering down-rift. Their injection sets up localized stresses in the central region that are relieved by the formation of intrusions approximately orthogonal to the general trend. Asymmetric addition of dykes to one side of a rift zone moreover can cause originally collinear rift zones to become non-collinear.

Propagation of bladed dykes

The conditions for lateral propagation of bladed dykes were comprehensively reviewed by Rubin & Pollard (1987). Lister & Kerr (1990) studied the fluid-mechanical aspects including effects of injection along the level of neutral buoyancy, and concluded that the resistance to dyke propagation by fracturing of crustal rocks is negligible in relation to that due to viscous pressure drop.

Lateral dyke-injection from a high-level magma chamber is thought to be important near volcanic centres, but further away dykes may commonly be injected vertically upward from a deep reservoir (Gautneb & Gudmundsson, 1992).

Magma-flow direction

In eroded volcanoes, the magma-flow direction in dykes can be inferred from the orientation of various structural features: shallow grooves on the dyke walls (Walker 1987), the elongation direction of dyke segments and 'fingers' at the leading edge of dykes (Pollard et al. 1975; Baer & Reches 1987), the crystal fabric of the dyke-rock (Shelley 1985), the deformation pattern of vesicles (Coward 1980), and the direction of pipe vesicles in dyke margins. Some, such as the last-mentioned feature, permit the absolute direction (azimuth) of flow to be inferred.

A powerful technique for inferring the flow direction is based on the magnetic fabric. Magnetic susceptibility, a measure of the ease with which a rock may be magnetized when exposed to a magnetic field, has an anisotropy that is thought to be related to a preferred orientation of elongated magnetic crystals. The maximum-susceptibility direction marks the magma-flow direction. Near dyke margins the maximum-anisotropy axis is not parallel with the dyke walls but has an imbricate-like orientation from which the flow azimuth can be inferred.

Knight & Walker (1988) found a wide scatter of flow azimuths in dykes of the Koolau complex, with an average azimuth upward at about 30° from the horizontal. Staudigel et al. (1992) found a similar variability in the Troodos sheeted-dyke complex. Ernst & Baragar (1992) found in dykes of the continent-ranging Mackenzie swarm in the Canadian Shield, that the magma-flow direction is vertical within about 500 km of the focus and horizontal farther out. Wada (1992) found in the 1983 dyke on Miyakejima that lateral magma flow was followed by near-vertical flow.

Re-injection intrusions

Back-draining often occurs in lava eruptions in Hawaii. Up to 2.2 million $m^3 h^{-1}$ of lava drained back into the Kilauea Iki vent during pauses in the 1959 eruption, for example (Macdonald 1962). Re-injection is easily explained because the neutral-buoyancy level for non-vesicular magma is well below the surface (about 1 km deep?) and if newly erupted vesicular lava loses bubbles its effective density rises and it may then descend to a neutral-buoyancy level. Note that this re-injection is not the same as water draining down into a volcano. The water in doing so passively percolates down into any available voids whereas the re-injected lava has the power actively to open old fissures or create new ones.

Generally re-injection generates dykes; they look like normal dykes but have glassy margins that are depleted in sulphur, lost in the 1-atmosphere conditions of surface eruption (Easton & Lockwood 1983). Anisotropy of magnetic susceptibility or other flow-direction indicators would demonstrate that downward magma-movement has occurred.

A class of small intrusion recently recognized in Hawaii is a kind of sill that transgresses downward at about 10° across the countryrock lavas in an outward direction extending as far as 10 km from the volcanic or rift zone centre. Mostly these intrusions are non-vesicular and rich in olivine, giving them a high density of

around 3.1 Mg m^{-3} about 0.8 Mg m^{-3} greater than average basaltic lava. From what is known of the fracture toughness of lavas, gravity is perfectly capable of propagating such intrusions. On the island of Kauai these intrusions locally comprise a swarm having an intensity of about 10%.

Craters, calderas and pit craters

Craters and calderas are depressions, commonly deep and precipitous, that mark the eruptive vents of volcanoes. Those at or near the volcano summit may directly overlie the magma chamber, and for those such as Stromboli (Giberti *et al.* 1992) that are persistently active a direct connection exists from the chamber to the conduit and vent. The most noteworthy are those few volcanoes that sustain a long-lived lake of molten lava in their summit crater or caldera.

The craters of basaltic stratovolcanoes are in no way different from the craters of more silicic volcanoes. They have elevated rims largely of pyroclastic deposits and can have an impressive depth exceeding 500 m. Not all basaltic craters are circular. The explosively generated Chasm of the Tarawera 1886 fissure eruption in New Zealand is a nearly straight trench 8 km long and 300 m wide by, on average, 100 m deep.

Calderas are subsidence features bigger than craters (Fig. 4) and by convention exceed about one mile in diameter. Those in subduction-related settings originated in explosive eruptions that emitted large volumes of pyroclastic material. Collapse of part of the volcano then took place into the magma chamber. Masaya caldera (Nicaragua) is related to an extensive basaltic ignimbrite (Williams 1982) and appears to have this origin. Williams & Stoiber (1983) proposed 'Masaya-type' for this the mafic analogue of 'Krakatau-type' calderas. The impressive calderas of Ambrym (Vanuatu; 12 km in diameter; Monzier *et al.* 1991) and Niuafo'ou (near Tonga; Macdonald 1948) also appear to have this origin.

Opinions differ whether caldera width in general is comparable to the magma-chamber width. In examples where cauldron subsidence of a cylinder of rock within a ring fracture occurred, the widths must be comparable; in other examples subsidence occurred in a localized funnel-like central area and the caldera then widened by scarp retreat. The scarcity of mafic ring dykes in eroded volcanoes (Chapman 1966) suggests that the second alternative may be more generally applicable to basaltic volcanoes.

Calderas of hotspot or rift-related volcanoes are generally not accompanied by a significant amount of pyroclastic deposits and appear thus to be due to subsidence with little explosive activity.

Two basaltic caldera-collapse events are known to have occurred in historic time: the 2 km^3 Oskjuvatn caldera on Askja volcano in Iceland formed by subsidence in 1874–1875, and part of the floor of Fernandino caldera in the Galapagos subsided in 1968 to increase the caldera volume by 1–2 km^3.

The Oskjuvatn event accompanied rifting in which lateral flow of basalt magma evidently took place to Sveinagja, some 50–70 km north of Askja, where lava erupted. A short plinian eruption of rhyolitic pumice also occurred from a vent in Oskujuvatn. The 0.3 km^3 of erupted basalt plus the 0.2 km^3 (dense-rock equivalent volume) of rhyolitic pumice was only a fraction of the caldera volume, and the balance may be contained in the dyke that solidified in the fissure (Sigurdsson & Sparks 1978).

In the Fernandino event (Simkin & Howard 1970) a large part of the 7 km^2 floor of the already-existing caldera subsided within an elliptical area by as much as 300 m. The subsided block sagged slightly in the middle suggesting that the bounding faults had an inward dip. Explosions also occurred and a large ash plume rose, but the ash volume was minor compared with the subsided volume. A lava eruption occurred on the volcano flank but the lava-volume was also minor. Possibly a larger eruption took place unseen on the submerged flank.

Caldera collapse on basaltic volcanoes is not necessarily related to magma discharge: Walker (1988) proposed that Hawaiian calderas are funnel-like structures caused by downsagging, due to the weight of intrusions, into thermally weakened lithosphere. The inward dip shown by many layered gabbros may be symptomatic of central sagging.

Pit craters are similar to, but smaller than, calderas. Fine examples occur along the active rift zones on Kilauea and are ephemeral structures that appear, widen by wall collapse, grow by coalescing with other pit craters, and are destined eventually to be infilled with lava. The largest, Makaopuhi, is 1.6 km long by 1.0 km wide, and prior to partial infilling in 1965–74 was 300 m deep. The youngest is Devil's Throat, formed in about 1921. The caldera of Mauna Loa (called Mokuaweoweo; Macdonald 1965) and also that of Grand Comore (Strong & Jacquot 1971) originated in part by coalescing pit craters.
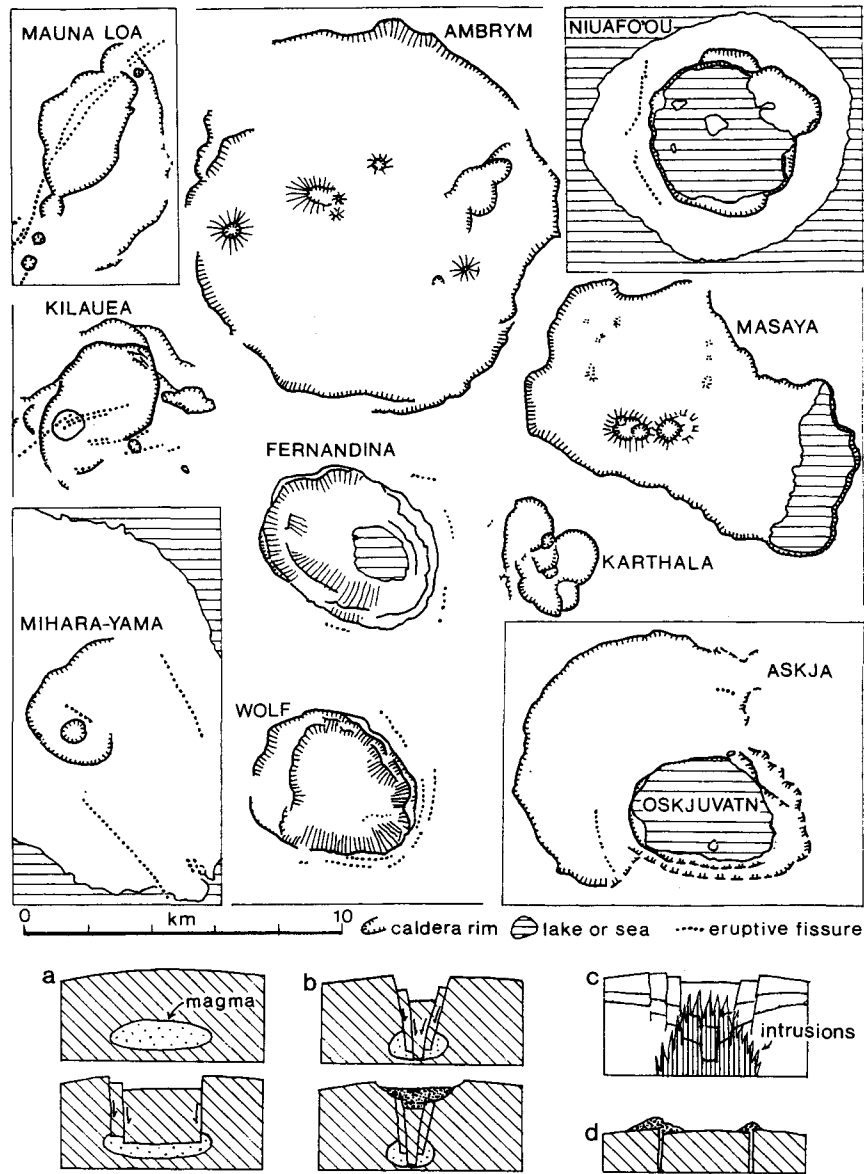
Walker (1988) proposed that pit craters of

Fig. 4. Craters, calderas and pit craters. (Above) Maps on the same scale of basaltic calderas; north is up; Mauna Loa and Kilauea in Hawaii, after Walker (1988); Mihara-Yama in Izo-Oshima, Japan; Ambrym in Vanuatu after Monzier *et al.* (1991); Niuafo'ou near Tonga, after Macdonald (1948); Fernandina and Wolf in the Galapagos, after Munro (1992); Karthala in Grande Comore, after Strong & Jacquot (1971); Askja and Oskjuvatn in Iceland.

(Below) Schematic sections showing proposed caldera-forming mechanisms – (a): subsided cylinder – cauldron subsidence into magma chamber; (b): narrow pit caused by subsidence into magma chamber, widened by mass wasting; (c): funnel-like downsag caused by load of intrusions (Walker 1988); (d): ring mountains generated by eruptions on annular fissures (Brown *et al.* 1991).

Kilauea are surface manifestations of a deep sub-horizontal duct that conveys magma intermittently from the summit chamber into a rift zone. In an active duct, localized roof collapse may generate vaults that extend upward by roof collapse, some of the debris being carried away by lava, and eventually the vault breaks surface.

Some pit craters contain long-lived lava lakes, notably those of Erta'ale (Ethiopia; Le Guern *et al.* 1978), Nyiragongo (Zaire; Tazieff 1977, 1984) and Kilauea prior to 1924 (Macdonald *et al.* 1983). Their depth varies as periods of infilling by lava alternate with subsidence or drainback events.

Particularly deep drainback allows hot groundwater to escape and permits groundwater to enter the hot conduit, leading to violent phreatic steam explosions such as those of 1790 and 1924 on Kilauea (Decker & Christiansen 1984). A catastrophic escape of lava from the Nyiragongo lake in 1977 produced an exceptionally fast-moving lava flow on the volcano flanks that overwhelmed many people and elephants (Tazieff 1977).

Magma chambers, neutral buoyancy, and cumulate prisms

Probably all polygenetic volcanoes at or near the peak of their activity possess a magma chamber. This chamber is more than simply an underground pool of magma: it occupies a central position in the volcano, it serves as a storage container in which magma newly arrived from the mantle accumulates, and it modulates the supply of magma to intrusions and to the surface. Fractionation occurs in it to generate on the one hand more evolved magmas and on the other great masses of mafic to ultramafic cumulate rocks. Processes that operate in magma chambers are reviewed by Marsh (1989).

The depth of Kilauea's magma chamber, as inferred from ground-deformation data and the position of an aseismic zone thought to be occupied by magma, is between 0.6 and 9 km (Ryan 1987; Ryan *et al.* 1981). Seismic probing is more favourable under water and reflection techniques have identified reflectors, regarded as magma, at a depth of 2–3 km below sea-bottom in the East Pacific Rise (Detrick *et al.* 1987). These chambers appear to be strongly elongated parallel with the ridge axis, are several kilometres wide, and are thin.

Long-lived magma chambers occur in neutral buoyancy positions where, in a vertical sense, they are in gravitational equilibrium. A level of neutral buoyancy is normally, however, a laterally extensive surface, and gravitational forces tend to cause the chamber to expand laterally along this surface. Expansion is opposed by the strength of wall-rocks and cooling of magma as it extends into cold rocks.

For these reasons a magma chamber is in only a partial state of gravitational equilibrium. As it inflates due to the input of newly arrived magma or the formation of gas bubbles, so rupture of the chamber walls or roof eventually occurs and a magma excursion results.

Little is known about the sizes or shapes of magma chambers of subaerial basaltic volcanoes. Some, but not necessarily all, gabbro intrusions seen in eroded volcanoes are solidified magma chambers, but it is generally unlikely that the whole volume of the intrusion was fluid at one time. An example of a large intrusion that apparently is a solidified chamber is the Kap Edvard Holm gabbro in East Greenland, the minerals of which show that it behaved as an open system (Bernstein *et al.* 1992), whereas nearby Skaergaard intrusion that shows extreme fractionation behaved as a closed system (McBirney & Noyes 1979).

Fedotov (1982) investigated the growth and extinction conditions for magma chambers. He showed that they grow and reach a maximum and stable diameter of several kilometres, at which heat influx balances heat losses, within a few thousand years. In a similar study Hardee (1982) found that a narrow sill may become a confluent intrusion or magma chamber at a magma influx rate of $> 10^{-3} \, \text{km}^3 \, \text{a}^{-1}$.

Gudmundsson (1987) assumed that magma exhibits poroelastic behaviour (i.e. melt occurs in pore spaces of a brittle crystalline framework), and from compressibility considerations calculated from the size of Icelandic magma excursions, assumed that the volume of magma chambers is typically in the range 100–1000 km³, about 2000 times the excursion-volume.

If a magma chamber is enclosed by rocks that are elastic, when the chamber swells elastic strain energy is then stored in the magma. When an excursion occurs the magma-flux should decay exponentially. Records of some actual eruptions (Wadge 1981) conform poorly with an exponential decay. A possible explanation is that while eruption occurs, so a partial replenishment occurs from the deep magma-source.

Little is known of magma-chamber shape. A complex of interconnected dykes and sheets is envisaged by some (e.g. Fiske & Kinoshita 1969) but this is not a stable configuration. Incoming magma would likely enter the hottest and most fluid part of the chamber, and heating of reentrants and cooling of salients would tend to round the outlines. The most likely shape would

be that of a triaxial ellipsoid with the longest and intermediate axes lying in the neutral buoyancy plane, and the longest axis extending into rift zones. The ellipsoid might be deformed into a funnel-like shape by floor subsidence, reaching a maximum in the centre. A shape something like this is not unusual among gabbro intrusions: the hypersthene gabbro of centre 3 at Ardnamurchan (Wells 1954) and the Ben Buie gabbro in Mull (Lobjoit 1959) are examples. In some volcanoes the chamber may simply consist of the cylinder of magma underlying the summit crater.

Cumulate prisms

Magma chambers of large basaltic volcanoes are underlain by mafic and ultramafic plutonic rocks, such as gabbro, dunite and harzburgite. Many of these rocks are cumulates and originated by the sedimentation of pyrogenetic minerals from the magma. Various lines of evidence point to the existence of such prisms.

(1) Great cumulate prisms are exposed in the cores of some deeply eroded volcanoes notably those of Rhum (Emeleus 1987) and the Cuillin Hills of Skye (Wadsworth 1982).
(2) Inclusions of cumulate rocks are commonly brought to the surface by magma eruptions, particularly during declining phases in the life of the volcano when the high-level magma chamber has solidified (e.g. Hualalai volcano; Chen *et al.* 1992; Lesser Antilles volcanoes: Arculus & Wills 1980).
(3) Large positive Bouguer gravity anomalies are centred on the volcanoes (Fig. 5) and are greatest in the central area (e.g. in Hawaii; Strange *et al.* 1965; Kinoshito 1965).
(4) A 3 km deep drillhole into the positive gravity anomaly on Piton de la Fournaise (Reunion: Rancon *et al.* 1989) passed through gabbros and then cumulate rocks from 1.92 km to the bottom of the hole at 3 km depth.
(5) Seismic refraction lines across the centre of Koolau volcano revealed the existence of rocks with a P-wave velocity of 7.7 km s^{-1} (near that of upper mantle rocks) at a shallow (1 km) depth (Furumoto *et al.* 1965).

It is envisaged that as a volcano is built up, so the prism of cumulate rocks grows and the neutral buoyancy level and magma chamber are displaced upward. The high-level chamber, which initially may have developed at the base of, or below, the volcano, migrates upward into the volcano, while the dense cumulate rocks accentuate the zonation that governs neutral buoyancy positions.

The excess of mass indicated by each of the Rhum, Cuillins and Mull gravity anomalies is $3-7 \times 10^{11}$ tonnes and can be modelled by prisms of mafic and ultramafic rocks $0.2-0.3 \text{ Mg m}^{-3}$ denser than crystalline basement rocks, having volumes of $1100-3500 \text{ km}^3$ (McQuillin & Tuson 1963; Bott & Tuson 1973). The well-exposed prism of Rhum is 10 km in diameter at the top and might thus extend downward some 15 km. The diameter of the prism should be related to that of the chamber.

Ultramafic inclusions

Magmas very commonly bring inclusions (xenoliths) of coarsely crystalline mafic or ultramafic rock to the surface. These inclusions because of their high density (commonly $3.0-3.3 \text{ Mg m}^{-3}$) relative to basaltic magma (about 2.75 Mg m^{-3}) and common large size ($>0.1 \text{ m}$) have high settling rates through Newtonian magma ($>1 \text{ cm s}^{-1}$) for a viscosity of 10^3 Pa s; their appearance at the surface indicates either ascent velocities higher than this, or possession by the magma of a yield strength (Sparks *et al.* 1977*a*).

More importantly, as pointed out by Clague (1987), such inclusions drop out if the magma pauses or resides at a level above their source; the transport to the surface of dunite cumulates from below the high-level magma chamber indicates that the high-level chamber has solidified, and the transport of mantle lherzolites and garnet pyroxenites to the surface indicates that no crustal magma chamber or reservoir exists at any level above the mantle.

Confluent intrusions

These are intrusions that grow incrementally. If a dyke or intrusive sheet is injected into an earlier dyke or sheet while the middle of the earlier intrusion is still partially molten, one is not chilled significantly against and merges into the other, and an intrusion of confluent type results. By successive increments, the intrusion widens and cools more slowly to produce a more coarsely crystalline rock. A succession of small intrusions may, thus, build a gabbro intrusion, the incremental origin of which may be difficult to detect. The gabbro may, however, be seen to pass at its lateral extremities into a cluster of dykes or sheets. Some time constraints for the increments are considered by Hardee (1982).

Shadow zones in central volcanoes

A significant feature of large central volcanoes is that each possesses a 'shadow zone' within which
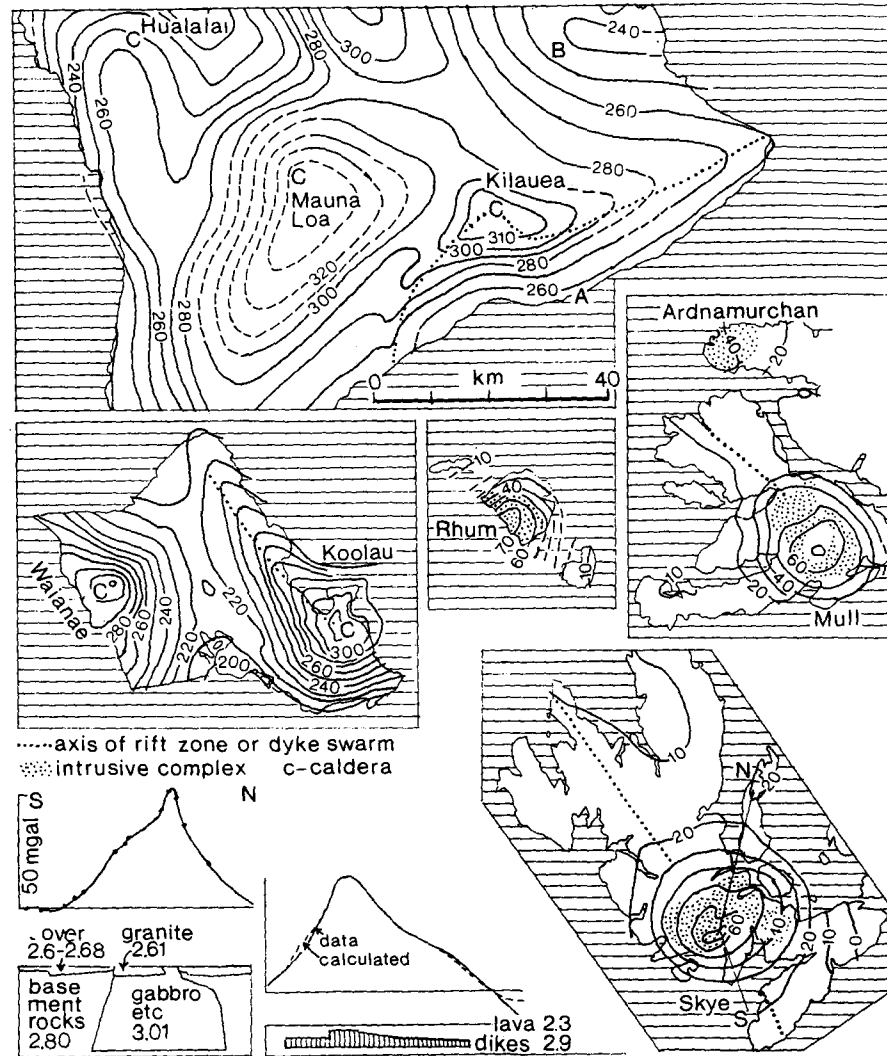
Fig. 5. Magma chambers, neutral buoyancy, and cumulate prisms. (Main picture) Bouguer gravity anomaly maps, on the same scale, contour interval 10 mgals, of Mauna Loa and Kilauea volcanoes, Hawaii (Kinoshito 1965); Koolau and Waianae volcanoes in Oahu (Strange *et al*. 1965); and intrusive volcanic centres of Rhum (McQuillin & Tuson 1963), Mull (Bott & Tuson 1973) and Skye (Bott & Tuson 1973). (Below left) gravity profiles across Skye and Kilauea, and density structure in the upper crust that is capable of producing the observed anomaly (Kilauea: after Furumoto 1978). Rock-density values in $Mg\,m^{-3}$.

basaltic vents are rare or absent. The shadow zone is interpreted to overlie a body of silicic magma through which ascending basaltic magmas are normally unable to pass (Hildreth 1981). Examples are Newberry, Pantelleria, and Torfajokull (Iceland; Walker 1989*a*).

Silicic vents are concentrated in the shadow zone. It is thought that basaltic magma incursions heat the silicic body, promote convection by reducing its viscosity and density and cause

vesiculation (the solubility of gas in magma decreases as the temperature increases). In effect, the basaltic magma 'stokes' the silicic magma and causes changes that favour or trigger its eruption (Sparks *et al*. 1977*b*).

Some basaltic magma batches may succeed in rising through the silicic magma body if their ascent rate is high or if the silicic magma has a high yield strength. They locally heat and mobilize the silicic magma, provide a hot pathway for

it toward the surface and thus play an active role in promoting silicic eruptions. This is thought to have happened in the composite dykes having basaltic margins and a silicic centre, and composite lava flows having a basaltic base overlain by rhyolite, that occur in Iceland and the Hebridean Province.

It has evidently also happened in the case of some silicic eruptions that were immediately preceded by a basaltic eruption. A fine example is the great Rotoehu Ash/Rotoiti Ignimbrite eruption from Okataina volcano, New Zealand, about 50 000 years ago which released some 100 km³ of rhyolitic magma and was immediately preceded by eruption of the volumetrically insignificant Matahi basalt.

Paradoxically, the more efficient the heat-exchanger system of a central volcano, the less clear is the evidence that basaltic magma participated. Thus, in the Taupo Zone a great volume of silicic volcanic rocks occurs but basaltic rocks comprise only about 1% of the total eruptive volume (Cole 1973). A few silicic units were erupted as mixed magmas (Blake *et al.* 1992), but generally contain only a very minor amount of basaltic material.

Net-veined intrusive complexes that are exposed in many deeply eroded central volcanoes (e.g. Southeast Iceland; Blake 1966; Mattson *et al.* 1986) provide some of the clearest evidence for injection of basaltic magma into silicic magma.

Major sill swarms

Sills underlie many flood-basalt fields. They can have large individual volumes, comparable with flood-basalt lava flows, and sill swarms can rival flood basalts in total volume (Walker & Poldervaart 1949). The sills mostly intrude thick sedimentary rock sequences that at the time of sill injection were young and poorly consolidated. One can infer that levels of neutral buoyancy existed between well- and less well-consolidated sediments, or alternatively the sill magma simply flowed under loose sediment. Bradley (1965) recognized that if a magma is denser than the rock it intrudes, the roof may effectively float on the magma. This, however, is an oversimplification of the situation since the magma has the capability of reaching the surface and building a volcano there (sills normally underlie volcanoes).

A swarm of Jurassic dolerite sills up to 300 m thick occurs in Tasmania and outcrops over 25% of the 65 000 km² surface area of the island. Parts of the same sill swarm are found on disrupted portions of Gondwanaland in Antarctica and South Africa (Frankel 1967). The Palisades and Gettysburg sills are of early Jurassic age and cut Triassic sediments under the Newark volcanics that extend from New Jersey to Newfoundland. Major sills including those of the Shiant Isles underlie the Tertiary flood basalts of Skye (Gibson & Jones 1991; Gibb & Henderson 1989).

Sills characteristically transgress across the countryrock bedding. Carey (1958) investigated this transgression in a Tasmanian sill by constructing isostrats – lines on a map connecting points where the sill is in contact with the same stratigraphic horizon – and demonstrated that the sill has the overall form of a saucer or low-angle inverted cone relative to the stratigraphy. He inferred that magma was injected at the low point of the cone. Francis (1982) found a similar pattern of isostrats in the Whin Sill and observed that the sill systematically thickens toward the low-point (Fig. 6a).

The Whin Sill, of late Carboniferous age, extends over at least 5 000 km² and has an average thickness of about 40 m. The aspect ratio is about 1/2000 and the volume is over 200 km³ (Francis 1982). The question arises, as with major flood-basalt flows, whether large-volume sills are emplaced rapidly or slowly. Gibb & Henderson (1992) and Husch (1990) present evidence that some sills form incrementally from successive magma incursions.

The injection of sills into poorly consolidated sediments commonly generates soft-sediment deformation structures in the countryrocks (Fig. 6g). Fine examples are described by Duffield *et al.* (1986). Einsele *et al.* (1980) pointed out that the volume contraction in baked and dewatered sediments may be sufficient to make space for a sill.

Peperites are rocks that are formed by mixing of magma with wet sediment, commonly in a weakly explosive manner. Peperites formed at margins of sills are described by Busby-Spera & White (1987) and Walker & Francis (1987), and very similar examples related to lava flows are described by Schmincke (1967). Yagi (1969) described a sill emplaced in wet sediment that partly subdivided into pillows.

Miocene flows and associated intrusions that outcrop along the coast of Washington and Oregon have been identified as distal portions of the Columbia River basalts that invaded, and locally flowed as shallow sills under, soft estuarine and deltaic sediments (Beeson *et al.* 1979; Pfaff & Beeson 1989).

It is probable that sills can form without surface volcanism. A possible example may be the sills penetrated by drillholes in the Fastnet
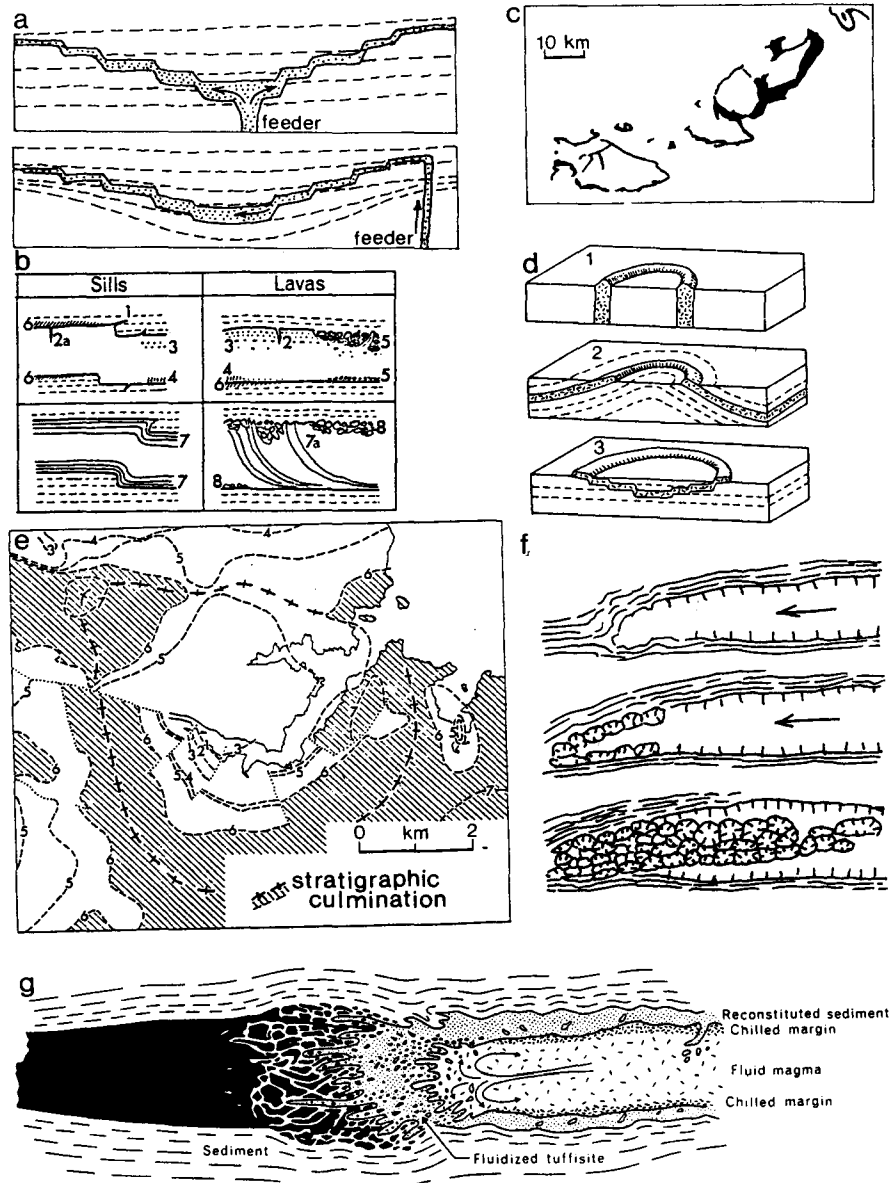
Fig. 6. Major sill swarms.

(a) Saucer-shaped sills; dashed lines are countryrock stratigraphic horizons; (upper) magma injected at the low-point (Carey 1958); (lower) magma injected at a high level and accumulating at bottom of sedimentary basin (Francis 1982). (b) Distinction between sills and lava flows: (above) more fluid magmas; (below) less fluid magmas. 1: cross-cutting of roof rocks; 2: infilled cracks; 3: vesicles; 4: pipe vesicles; 5: plant molds; 6: baked rocks; 7: flow planes; 7a: ramp structure of flow planes. (c) Ring-shaped outcrop pattern of the Gettysburg sill in the Triassic Newark Basin in Pennsylvania (Hotz 1952). (d) Intrusions that can give ring-like outcrops include: 1: Ring dykes; 2: folded sills; 3: saucer-shaped transgressive sills. (e) Isostrat map of a sill in Tasmania (Carey 1958). Dashed lines are isostrats for successively higher stratigraphic horizons 1 to 6. Shaded areas: sill in contact with rocks above horizon 6. (f) Inferred mechanism of formation of a pillowed sill (Yagi 1969). (g) Interaction of magma with coal (black) and unconsolidated sediments. (Walker & Francis 1987); reproduced by permission of the Royal Society of Edinburgh and authors from *Transactions of the Royal Society of Edinburgh: Earth Sciences*, **77** (1986), 295–307.

Basin (Caston *et al.* 1981) where coeval volcanism appears to be absent. The extensive sills in Tasmania are not accompanied by volcanism on that island although the Stormberg Series flood basalts are coeval with them.

Eruptions and eruption products

Basaltic lava flows

Lava flows are generated in most basaltic eruptions and form on about 20 volcanoes per year. They range over more than six orders of magnitude in volume and three in length, and generally have a low aspect ratio of <0.01. Many are of clastogenetic origin: the lava passes briefly through a fragmental stage in 'fire fountains' before the fragments coalesce and flow away. All gradations are found between such lavas and welded tuffs or agglutinate in which the distance flowed was insufficient to destroy the fragmental texture.

Close-proximal lava flows tend to be highly vesicular. In shelly pahoehoe coalescence and expansion of bubbles under the chilled skin generates hollow lava flows like gas bags with a roof crust commonly <0.1 m thick (Swanson 1973). Walking across a field of shelly pahoehoe can be quite hazardous where the crust is too weak to support a person's weight.

Pahoehoe and aa

Basaltic lava flows form two main structural types, one smooth-surfaced and called pahoehoe and the other rough-surfaced and called aa, which are fundamentally different in their structures and flow dynamics. As lava flows away from the source vents, the volumetric flow rate determines whether it becomes pahoehoe or aa (Rowland & Walker 1990).

At a low volumetric rate ($<20 \, m^3 \, s^{-1}$ in Hawaii) restriction of flow by rapid growth of a chilled crust causes the lava to be subdivided into small flow units averaging a few cubic metres in volume. In consequence the lava inside each unit (except for that which continues to flow under the crust) becomes static before cooling significantly.

In contrast, lava erupted at a high volumetric rate ($> 20 \, m^3 \, s^{-1}$ in Hawaii), destined to become aa, travels fast in open channels, and continues to flow after it has cooled significantly. Then, if the surface crust is torn by differential flow, the viscosity and yield strength of underlying lava are too high for it to well up and repair the tear. So by repeatedly tearing the lava comes to have a surface layer of spinose clinkery rubble frag-

ments and has become aa. 'Fingers' of lava that rise into the rubble carapace break on cooling joints and contribute debris to it. Morphometric features of aa flow-fields are described by Guest *et al.* (1987), Kilburn & Lopez (1991) and Wadge & Lopez (1991).

Aa flows invariably have an internal portion consisting of coherent basalt that underlies the surface rubble, and the coherent part averages about 60 per cent of the total flow thickness. Flowage has sometimes been likened to the advance of a caterpillar-track vehicle but this is a poor analogy since the rubble layer at the base is almost invariably much thinner than that at the top and is not simply the top rubble layer translated over the flowfront to the base. Over much of their travel distance aa flows possess a yield strength and move by plug flow. They effectively bulldoze obstacles in their path.

Traced from proximal to distal end, aa flows become progressively thicker and flowfronts commonly exceed 10 m high, while vesicles are strongly deformed and progressively eliminated until the lava becomes almost totally non-vesicular (except for new voids that are generated by shearing).

Pahoehoe flows advance slowly and the flow front consists of a multitude of small flow units which surge forward, become static, and spawn new flow units. Jones (1968) referred to this manner of advance as digital, referring to the numerous finger-like units. The flow units are fed through a system of lava tubes. In tubes the lava flows freely and without cooling significantly so enabling it to travel great distances (the Undara pahoehoe flow in Queensland is 160 km long; Stephenson & Griffin 1975) despite the low discharge-rate conditions.

Pahoehoe on slopes exceeding about 4° in Hawaii resides for only a short time in tubes and is highly vesicular (spongy or S-type pahoehoe; Walker 1989b). Pahoehoe on shallower slopes, as on the coastal terrace of an island volcano, resides longer in tubes, emerges with fewer but larger vesicles, and develops the highly distinctive pipe vesicles along the flow base (pipe-bearing or P-type pahoehoe; Wilmoth & Walker 1993). Pipe vesicles may be attributed to bubbles rising at the same rate as a rheology front. Behind this front the lava possesses a yield strength which prevents closure of lava behind the ascending bubbles.

An important aspect of pahoehoe flows particularly on shallow (<4°) slopes is the injection of lava under a surface crust, so jacking up this crust to produce lava rises or the localized whaleback-shaped uplifts, gashed by deep clefts, called tumuli. By this lava-rise mechanism a lava

flow initially < 1 m thick can thicken to > 5 m (Walker 1991; Hon *et al.* in press). Evidence is accumulating that some of the lava flows in flood basalt fields that were formerly thought to be erupted at extremely high discharge rates (Swanson *et al.* 1975) may in fact be lava rises formed by long-sustained and relatively low-discharge rate activity.

Pahoehoe lava shows a strong propensity to construct localized lava shields of scutulum type. A small-scale example is 110 m high Mauna Ulu, formed in the 1969–1974 eruption of Kilauea volcano. Much larger examples, exemplified by the 600 m high shield of Skjaldbreidur, occur in Iceland.

Some basaltic lava flows in subduction-related settings have a distinctly higher viscosity than those of other settings. This is inferred, for example, from the great flow-thickness, up to 30 m, of the flows erupted in 1759–1774 on El Jorullo (Mexico; Segerstrom 1950). Such lava may possess a yield strength at the time of eruption. This is inferred from the vesicle distribution pattern and occurrence of megavesicles in the Xitle basalt (Mexico City; Walker in press). Jorullo and Xitle lavas both erupted with a high content of crystals consistent with possession of a yield strength.

The surface aspect of these flows may be that of a block lava such as is more characteristic of andesite and silicic lavas than aa. The blocks are on the whole larger than the rubble fragments of aa and are bounded mostly by smooth cooling-crack surfaces.

Explosive eruptions and their products

Shield volcanoes produce a minimum of explosive products and have an explosivity index (i.e. percentage of pyroclastic deposits among their total products) as low as 10%. In contrast, stratovolcanoes in subduction-related settings typically have an explosivity index exceeding 50%.

On shield volcanoes at the peak of activity the eruptions are predominantly of hawaiian style, in which 'fire-fountains' play to a height of mostly tens of metres. Because of the large size of most lava fragments, nearly complete phase separation occurs so that only the lightest particles enter the soaring gas plume. Spatter ramparts are built along eruptive fissures, and cones or rings of spatter and cinders are built at points where effusion is concentrated. Most of the spatter lumps thrown up in eruptions of this style weld together when they land to form agglutinate or coalesce and flow away as lava flows.

Strombolian-style activity predominates on stratovolcanoes and monogenetic volcanoes.

The typical products are cinder cones. Cones occur in great numbers on the flanks of large stratovolcanoes; 200 occur on Etna and 60 on Fuji, and 132 occur on Mauna Kea shield volcano as products of its declining phases of activity. Cinder cones built in single eruptions range up to about 400 m high. They commonly include welded layers and contain variable proportions of spatter. The morphology of cinder cones is described by Wood (1980), and the dynamics of hawaiian- and strombolian-style activity are analysed by Wilson & Head (1981) and Head & Wilson (1989).

Cinder cones > 400 m high are known but are not called cinder cones. Usually they are stratovolcanoes buttressed by lava flows. An example is Izalco in El Salvador which was born 1769–1770 and has since grown to > 650 m high.

Strombolian cones often contain volcanic bombs. Some bombs are cored (Brady & Webb 1943). The core may be a piece of scoria or a mantle xenolith and is coated with basalt. Other bombs are fusiform (spindle shaped) and made of very dense material that evidently resided sufficiently long in the vent to lose most of its gas bubbles. It was formerly thought that the spindle shape is due to aerodynamic drag while spinning through the air, but Tsuya (1941) showed that it is produced by stretching, leading to necking and disruption of lava strands as they rose from the vent. Francis (1973) described cannonball bombs rounded by attrition as they bounded down the side of a cone, and Walker (in prep.) found potato-shaped lapilli and bombs 2–200 m in size on Tantalus, Oahu, rounded by tossing of still-plastic lava among loose ash in the vent.

Cinder-cone slopes are at the angle of repose of loose cinders (33–35° from the horizontal). Extensive redistribution of cinders by grain-flow occurs as the cone becomes oversteepened by excessive deposition immediately around the crater. The grain-flow deposits are recognized by their inverse size-grading and laterally-discontinuous lenticular bedforms. Many of the larger cinder fragments fall apart along cooling joints while in the grain flows. After ejection, the scoria vesiculates and develops a concentric vesicle zonation, and this zonation is truncated by the cooling joints.

In the most powerful strombolian activity, fire-fountains exceed 1 km high. They were up to 1600 m in the November 1986 eruption of Izu-Oshima (Aramaki *et al.* 1988). The brief October 1983 eruption of Miyakejima was particularly intense: the eruption lasted under 15 h at an average discharge of 1300 m³ s⁻¹ (Aramaki *et al.* 1986). In such intense activity, phase separation is incomplete, much of the ejecta enters the

convective plume, and an extensive blanket of fall deposits extends around the cinder cone.

Much of the explosive activity in subduction zone settings is better characterized as violent strombolian or vulcanian. The Paricutin eruption of 1943–1952 is a good example (of basaltic andesite, not basalt). The ejecta are highly fragmented and eruptive columns are dark coloured because of the abundance of ash. The ejecta tend to be dense and poorly vesiculated, and rich in lithics. It is inferred that the erupted magma had a yield strength that inhibited escape of gases and enhanced explosive violence; the generally crystal-rich condition of the magma is consistent with this inference.

The high degree of fragmentation and high density of violent strombolian or vulcanian ejecta favour the generation of pyroclastic flows. An extensive basaltic ignimbrite occurs at Masaya volcano (Williams 1982), basaltic pyroclastic flows were observed on Lopevi, Vanuatu, in 1960 (Williams & Curtis 1964) and Ulawun, Papua New Guinea, in 1978 (McKee *et al.* 1981), and pyroclastic flow deposits are found among the 1759–1774 ejecta in El Jorullo.

Phreatomagmatic eruptions

When copious amounts of water enter an erupting vent to mingle with the upjetting lava, a great column of steam is generated and the lava is quenched and highly fragmented (Wohletz 1986). Fall-out of ash builds an ash ring that within a few years may become lithified to a tuff-ring. The crater rim diameter of 0.5–1.5 km is significantly larger than that of strombolian-style cinder cones.

The resulting eruptive style is called surtseyan (after the Surtsey eruption in the ocean off Iceland, 1963–1965; Thorarinsson *et al.* 1964). Surtseyan eruptions occur frequently near the coast in shallow seas (140 m deep at Surtsey) or lakes. They belong to a class of volcanic eruptions known generally as phreatomagmatic or hydrovolcanic. Kokelaar (1983, 1986) and Sohn & Chough (1989) discuss significant features of such activity and its products.

When an eruption occurs on low-lying land where the rocks are permeable and the water table is high, the explosions excavate a crater extending well below the general land surface. A lake accumulates in the crater. If the erupted volume is small and the crater is surrounded by a low pyroclastic ring, the depression may be a more striking topographic feature than the pyroclastic ring. The resulting structure is called a maar (Lorenz, 1985, 1986).

Phreatomagmatic activity generates base surges, recorded by the dune-bedded structures of their deposits. The type examples of base surges formed in the 1965 eruption of Taal volcano (Philippines; Moore 1967). These surges swept outward to 4 or 5 km from the eruption crater. Trees that remained standing were coated with mud and ash on their up-vent side. The deposits on the ground had a wavy surface consisting of ring dunes tangential to the crater, showing internal dune bedding (Fisher & Waters 1970). Great numbers – scores to hundreds – of base surges are recorded in some tuff-rings.

Some eruptions are comparatively wet: their ash deposits contain accretionary and ash-coated lapilli, show intra-formational gullying and slumping, and pitting of some bedding planes by rain-drop impressions. Others are relatively dry, have a lower content of fine ash, and mark a transition to strombolian-style activity (Verwoerd & Chevallier 1987).

Phreatomagmatic vents quarry down often some hundreds of metres into the underlying rocks, and pieces of these rocks are abundant as included lithics in the pyroclastic deposits. Blocks also occur deeper than the level from which they originated. In ancient examples where the surface deposits have been eroded away the vent survives as a diatreme, a more or less cylindrical neck or pipe infilled with pyroclastic material.

Rootless vents

Steam explosions commonly form craters and pyroclastic accumulations where lava flows from land into water. These structures were referred to by Thorarinsson (1953) as 'pseudocraters', but the craters are real and are better referred to as rootless vents. At the type locality in Iceland where a lava invaded the shallow Lake Myvatn, many hundreds of pyroclastic structures occur ranging from hornitos little more than 1 m across to ash-rings of 0.5 km rim diameter.

Littoral cones are common in Hawaii at the point of entry of lava into the sea. The most common type consists of a pair of half-cones on either side of the lava (pyroclasts that fell onto the lava were carried away), exemplified by Puu Hou (Fisher 1968). These pyroclastic accumulations appear identical to those at primary vents except that the density of the pyroclasts is commonly higher (reflecting gas losses from the flowing lava; Walker 1992b). A newly recognized type of littoral cone, exemplified by Puu Ki, that occurs on some pahoehoe flows of Mauna Loa, consists of complete and nested craters about a central rootless vent apparently fed from a lava tube (Jurado *et al.* in prep.).

Phreatic eruptions

These are steam explosions in which no juvenile magma is erupted. Some kind of heat-exchange mechanism operates in which magmatic heat is transferred to water of external origin to power the explosions.

Underwater basaltic volcanism

A large proportion, probably amounting to $3.5 \, \text{km}^3 \, \text{a}^{-1}$, of the world's basalt is erupted under water, mostly at divergent plate boundaries (spreading ridges). Knowledge of this underwater volcanism is derived from bottom studies (including observations from submersibles), sea-floor drilling, and investigation of ophiolites that appear to be tectonically uplifted portions of oceanic crust. Valuable supplementary information is supplied from places such as Iceland, where widespread volcanism occurred under the extensive ice sheets (now largely melted) of the Ice Age.

The principal difference between underwater and subaerial volcanism is that explosive volcanism is suppressed under water. The deeper the water, the more strongly it is suppressed. This is because, under the confining pressure of water, less volatiles are exsolved from magma than at the surface, and also because a given mass of exsolved gas has a much smaller volume. Under a confining pressure of 0.1 kb appropriate to a water depth of 1 km, for example, the solubility of water in basalt is 1.2 wt% (cf. <0.1% at 1 atmosphere pressure), and 1 g of exsolved water vapour at 1150°C has a volume of only 0.06 l (cf. 6 l at 1 atmosphere). Explosive fragmentation of magma requires that the volume fraction of gas bubbles should exceed about 0.7, likely to be rare except in water less than about 200 m deep.

Pillowed lava

Many of the lava flows erupted under water are compound and subdivided into small rounded flow units called pillows. Pillows are very similar in size and shape to pahoehoe flow units but tend to be narrower, have a larger aspect ratio (vertical dimension/horizontal dimension), and have continuous glassy rims. Pillow lava seen in cross-section has an appearance like that of a pile of sandbags, and formerly it was considered that each pillow is a discrete sack-like lava body. It is now thought that generally pillows are connected by narrow tubes that are intertwined rather like spaghetti (Jones 1968).

Conceptions of pillow-lava growth were strongly influenced by direct observations made by diving on the 1969–71 lava delta of Kilauea volcano (Moore 1975). In this instance the pillows flowed down an underwater slope of 20° or more and were strongly elongate downslope. Pillows have not yet been observed forming on shallow slopes.

Pillow lavas are remarkably similar to pahoehoe, but are on average less vesicular. Pipe vesicles are common in pillows on small-angle slopes. Tumuli and tumulus-like features from which many pillows have issued have been described by Appelgate & Embley (1992) and Ballard & Moore (1977). Sheet lavas also occur (Ballard et al. 1979) and form at a higher discharge rate (cf. Griffiths & Fink 1992). Commonly sheet lava has a pillowed base or top. Tribble (1991) observed lava flowing in open channels, and the ultrathin lava layers described by Moore & Charlton (1984) may be similar, formed in channels or as multiple crusts in lava tubes. Even where vesicles are scarce, large voids can occur in pillows at all depths due to drainage of lava from tubes (Fornari 1986).

Hyaloclastite lava

Another kind of lava flow that forms under water may be designated a hyaloclastite lava: like aa it has a fragmental carapace and a coherent core. The fragmented top layer tends to be thicker and finer grained than in aa, and the fragments tend to be less vesicular and more glassy, smooth surfaced and angular in shape, due to their fragmentation by the shattering and crumbling of lava in contact with water.

Rittmann (1962) introduced 'Hyaloclastite' for rocks fragmented by the quenching and granulation of hot lava in water, and 'hyaloclastic' for the process of producing a hyaloclastite (Greek hyalos = glass, klastos = broken). Not all 'broken glass' rocks are included: only those broken in contact with water. Note that the products of surtseyan eruptions are also glassy but some at least of the juvenile ejecta are highly vesicular. They are strictly pyroclastic and are better called 'hyalotuffs' (Honnorez & Kirst 1975).

A hyaloclastite lava occupying a glacial valley was described by Walker & Blake (1966) in southeastern Iceland. By inversion of topography it now constitutes a ridge (called Dalsheidi). The valley was occupied by a glacier when Dalsheidi erupted. A highly irregular body of coherent basalt that occurs near the bottom of the palaeovalley marks the master conduit through which lava flowed down-valley. It shows conspicuous prismatic jointing everywhere.

The hyaloclastite carapace of Dalsheidi, up to

300 m deep, is envisaged to have grown endoge-nously as lava escaped from the conduit, and a system of irregular dykes and intrusive sheets that branch off from the main conduit represent the pathways taken by this lava. Such dykes that occur within and are a part of a lava flow are called lava dykes (Silvestri 1962). Lava dykes commonly occur also in pahoehoe compound flows. The most important features of Dalsheidi were the evidence it provided that lava is capable of flowing for tens of kilometres below ice (and by implication under water) and that a hyalo-clastite carapace may grow endogenously.

Much more extensive hyaloclastite lavas are described from southern Iceland by Bergh & Sigvaldason (1991). They form sheets up to 200 m thick, and individual flows cover up to 280 km². Traced laterally a number of distinct facies are observed including mass-flow hyalo-clastites.

A particularly favourable environment for hyaloclastite formation is the littoral zone where lava enters water from land, and where the repeated dashing of waves against freshly exposed incandescent lava causes rapid and voluminous fragmentation. Generally the lava and hence the hyaloclastic fragments have a low to moderate content of vesicles.

Lava deltas and tuyas

Pahoehoe lava that flows into the sea extends the shoreline outward as a lava delta (Jones 1969; Moore et al. 1973). An abrupt passage zone separates subaerial lava of the delta from the underlying flow-foot deposits formed under-water. Flow-foot rocks are mainly hyaloclastic breccias containing pillows. Commonly they dip at 20–25° like the foreset beds of a river delta, and the pillows are strongly elongated in the dip direction. Furnes & Fridleifsson (1974) described regular fluctuations in level of the passage zone in an Icelandic lava delta and attributed them to the tidal variations in water level; since each tidal cycle is about 13 hours, this provides a means of calculating the rate of ad-vance of the delta.

Certain table mountains called tuyas, com-mon in Iceland and known also in British Col-umbia, Alaska and Antarctica, are distinctively-shaped volcanoes formed by eruptions under former ice-sheets. Each tuya has a core and pedestal of pillow lava and hyaloclastite overlain by hyalotuff. This is capped by a lava-delta that formed in an intraglacial meltwater lake, and the relatively flat top consists of subaerial lavas of the delta (Mathews 1947).

Tuyas in Iceland are mostly monogenetic

structures and are the intraglacial equivalent of scutulum-type shields. They commonly have a volume of between 1 and 10 km³. Tuya-like structures also form by eruptions in the sea: Surtsey, which consists of hyalotuffs capped by a lava delta, is an example. Overlapping tuyas may have passage zones at widely different levels related to different thicknesses of ice sheet or levels of the sea (Jones & Nelson 1970).

Aa lava that flows into the sea tends to pro-ceed as though the water is not there, and may have an underwater flow width and flow thick-ness little different from those on land. The extension of the coastline is minimal (Moore et al. 1973). One can infer that distal-type aa exhibits this behaviour because it possesses a yield strength.

Palagonite and palagonitization

Glassy basalt readily hydrates to palagonite, an orange-coloured hydrogel and its fibro-crystal-line derivative. Chemical constituents that are leached out crystallize commonly as zeolites in void spaces and effectively lithify the loose de-posits (Furnes 1974). Palagonite forms rapidly under hydrothermal conditions, and its rate of formation doubles with every 12°C temperature increase. When a drillhole passed through the Surtsey hyalotuff 12 years after the eruption virtually the entire deposit was well lithified (Jakobsson & Moore 1986).

Jointing in basaltic rocks

Cracks (joints) are ubiquitous in basaltic rocks but few systematic studies of them have hitherto been attempted. They have not been employed to any significant extent as a tool from which to infer processes and conditions, and no review exists.

Contraction joints

When a volcanic rock body such as a lava flow or dyke cools from magmatic to atmospheric temperature it undergoes a volume contraction by several percent. This contraction is accommo-dated in part by the formation of contraction joints which develop at right angles to the cool-ing surfaces and subdivide the rock body into crude prisms. Contraction joints are more closely spaced near the cooling surface and the spacing widens further where the cooling rate is lower. Commonly a parting occurs in or near the middle of the body where joints that propa-gated inward from one cooling-surface of the body meet joints that propagated inward from
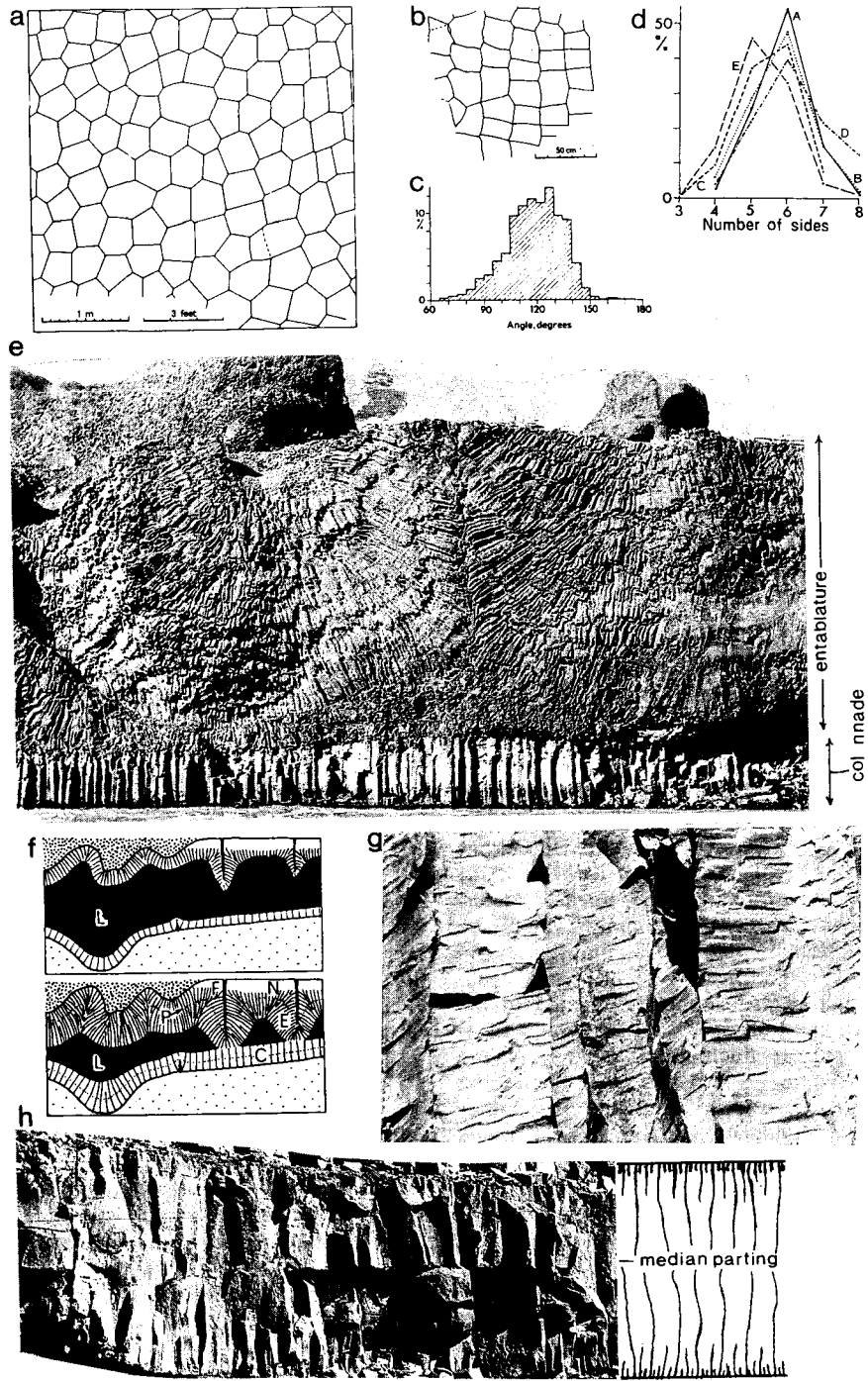
the opposite cooling surface. The pattern of jointing is one of the basic criteria for delineating cooling units within volcanic rock-bodies (Fig. 7h).

Prismatic joints commonly have band-like markings (called chisel structure by James 1920) on their surface. The bands are typically a few centimetres wide and approximately parallel with the cooling surface. Ryan & Sammis (1978) showed that chisel marks are due to incremental crack-propagation into the cooling rock body. They called them 'striations' (Fig. 7).

De Graff et al. (1989) showed how the joint-propagation direction can be inferred from surface markings on joint surfaces. Peck & Mina-kami (1968) observed that the incremental joint propagation can be detected in a cooling lava-lake by seismometer.

Jointing that subdivides the rock into prisms that are unusually regular in form and uniform in size is called columnar. In lava flows it is best developed where the lava occurs in valleys and topographic depressions. From this it is inferred to be favoured by cooling in static conditions, so that stresses resulting from volume contraction are uniformly distributed. Where a lava flow overlies water-rich sediment such as lignite, dew-atering is accompanied by a volume contraction of the sediment that helps generate the topographic depression occupied by the lava.

Columnar-jointed lava flows commonly show a multi-tiered structure in which broad and regular columns form a lower part (called the colonnade; Tomkeieff 1940) and a zone of narrow and irregular curvicolumnar prisms forms an upper part called the entablature. Typically there is a very abrupt transition from colonnade to entablature. Sometimes the entablature is capped by an upper colonnade (Fig. 7e).

Saemundsson (1970) proposed that the entablature is caused by water cooling, particularly where it is much thicker than the colonnade. Supporting evidence for water cooling is that the glassy mesostasis is more abundant in the entablature-lava than in the colonnade-lava (Long & Wood 1986). Lava flows occupying river valleys are particularly liable to be water-cooled because

the displaced river flows over the top surface of the lava. The upper colonnade formed before water cooling began.

Not all contraction joints in volcanic rocks are related to cooling. Palagonite-rich tuffs commonly exhibit prisms several millimetres to several centimetres wide, attributed to a volume reduction due to a partial desiccation of the palagonite when exposed to the air. The prisms are orientated normal to today's land surface showing that they are not palaeostructures but are due to drying out in today's atmosphere.

Expansion (inflation) joints

An important class of joint in volcanic rocks has a significant gape and results from an expansion of a volcanic rock body causing disruption of the chilled crust. Familiar examples are the distinctively jointed breadcrust blocks so often thrown out by explosive eruptions of andesitic volcanoes and the distinctive whaleback hillocks called tumuli in pahoehoe flow-fields.

The gaping cracks in breadcrust blocks form as the centre of a block expands due to continued vesiculation (Walker 1968). Tumuli grow by the injection of lava under a solid crust, jacking up the crust (Walker 1991). They are deeply gashed by V-shaped lava-inflation clefts that grew at the same time as the lava crust, their tip projecting into red-hot lava. Tumuli are localized uplifts. Lava rises are more extensive uplifts and similar clefts occur around their margins.

Lava-inflation cracks form also in the underwater environment. Thus, tumuli occur in underwater flowfields (Appelgate & Embley 1992) and many pillows have gaping expansion cracks adjacent to which partial springing-off of the chilled selvage allows water to penetrate and generate multiple selvages. Alternatively multiple selvages are due to implosion (Yamagishi 1985).

The opening of cracks enables water to penetrate deeply into a cooling lava; when secondary joints then develop normal to these new cooling surfaces they generate the highly distinctive joint system of 'pseudo-pillow lava' (Watanabe &

Fig. 7. Jointing in basaltic rocks.
(a) Plan view of columnar joints in the Giant's Causeway basalt. (b) Plan view of a less regular pattern of cooling joints. (c) Histogram of angle in degrees between column sides at the Giant's Causeway (O'Reilly 1879). (d) Percentage of columns having a given number of sides, at the Giant's Causeway (A; measured by Walker) and several other columnar lavas (Beard 1959). (e) Two-tiered columnar jointing in the basalt at Hloduklettur, NE Iceland. The lava is ponded in a river valley, and the irregular top is due to emplacement below its own cinder accumulation. (f) Stages in the growth of (e). L: liquid lava; F: master joint allowing ingress of water coolant; C: colonnade; E: entablature. (g) Chisel structures several centimetres wide on column surfaces, Iceland. (h) Lava flow about 5 m thick cut through by the Dettifoss canyon in NE Iceland, showing distribution of cooling joints. The overlying and underlying flows have been cut off the photograph for clarity.

Katsui 1976). The same joint pattern commonly occurs in the entablature of columnar basalts. Such cobble-jointed rock is called 'kuppaberg' in Iceland.

Not all expansion joints in volcanic rocks form during cooling. Many basalts contain a glassy mesostasis, and this is particularly liable to hydrate to palagonite. Hydration accompanied by expansion may then cause additional joint systems to develop. The ball and socket joints that subdivide the columns of the Giant's Causeway (Preston 1930) may be such a system. Column rinds that form by weathering (Smedes & Lang 1955) cause tensional stresses in each column that are relieved by the formation of cross-joints.

Flow jointing

During flowage of lava flows or magmatic intrusions, shearing occurs that may cause platy crystals (e.g. of plagioclase feldspar) to be orientated into near-parallelism with one another (this parallelism is called a trachytic texture), or cause gas bubbles to be deformed and their planes of flattening or elongation to be similarly orientated. This gives the rock a foliation along which it may split readily into platy fragments, particularly when accentuated by weathering.

Commonly the foliation/flow jointing exhibits ramp structure, being mostly inclined inward and up-flow along planes that curve across the flow and in plan view are concave in the up-flow direction. In longitudinal section these planes are steep at the flow-top and curve down to asymptote against the flow-base in the up-flow direction.

Vesicularity of basaltic rocks

Gas bubbles or vesicles are ubiquitous in basaltic flows. They form mainly by exsolution of dissolved magmatic gases in the volcanic conduit under the vent where, by their expansion and the positive buoyancy that they confer on the magma, they are a powerful driving force for eruption. Water of external origin that enters the magmatic system may contribute to the gas budget at any stage. A massive loss of gas occurs at the vent, and as lava flows away it steadily becomes degassed. The final gas loss occurs as the lava crystallizes.

The pattern of degassing, and features of the vesicles that remain in lava flows, can yield important information on flow mechanisms and lava rheology. This brief chapter concentrates attention on this aspect of magmatic gases. It is brief because few studies have yet pursued this topic (Fig. 8).

A general relationship is that gas bubbles tend to rise and grow by coalescence (Sahagian et al. 1989) so increasing their ascent velocity. The lower and middle parts of a flow thus become depleted in vesicles, while vesicles become concentrated near the flow top (Aubele et al. 1988). In general, the thicker the lava flow and hence the longer it takes to solidify, the more nearly complete is the loss of bubbles from the lower and middle parts of the flow.

Vesicle shapes, sizes and distribution patterns are very sensitive to lava rheology. Deformed bubbles stay deformed if the combined effect of surface tension and gas pressure fails to overcome the yield strength. Possession by lava of a yield strength can prohibit the ascent of all bubbles smaller than a threshold size.

If a lava flow is initially Newtonian but a rheology front ascends into it from the flow-base, bubbles that ascend at the same rate as the front become pipe vesicles because their lower end fails to close. Pipe bubbles grow as they ascend by scavenging smaller bubbles, and may become megavesicles (exceptionally as much as 1 m in size) that ascend diapirically and tend to accumulate under the surface crust (Walker in press).

At a late stage, a residual low-melting-temperature fraction of melt and gases expelled by crystallization escape from the crystal mush leaving microscopic angular voids ('diktitaxitic' texture; Dickinson & Vigrass 1965). This fluid segregates, ascends in vesicle cylinders commonly 5–10 cm wide, and injects as near-horizontal and highly vesicular segregation veins near the middle of the lava flow where viscosity and yield strength begin to increase upward. Segregation veins propagate laterally by hydraulic fracturing. Partial separation of gas bubbles from the segregation melt generates segregation vesicles (Smith 1967) floored by segregation melt. The segregated rock is characteristically not chilled against the host basalt.

Thin (<2 m) flow units of spongy ('S-type') pahoehoe that occur in Hawaii on groundslopes >4° are initially Newtonian, permitting larger mafic crystals to settle toward the base giving an S-shaped profile, but they cooled and acquired a yield strength before the initially small vesicles ascended or grew significantly. The vesicle-size and concentration curves therefore tend to be bilaterally symmetrical about the median plane giving a D-shaped profile (Walker 1989b). The vesicle concentration near the middle may be so high that a median parting has developed, the roof of it upbowed to form a gas blister. Gas
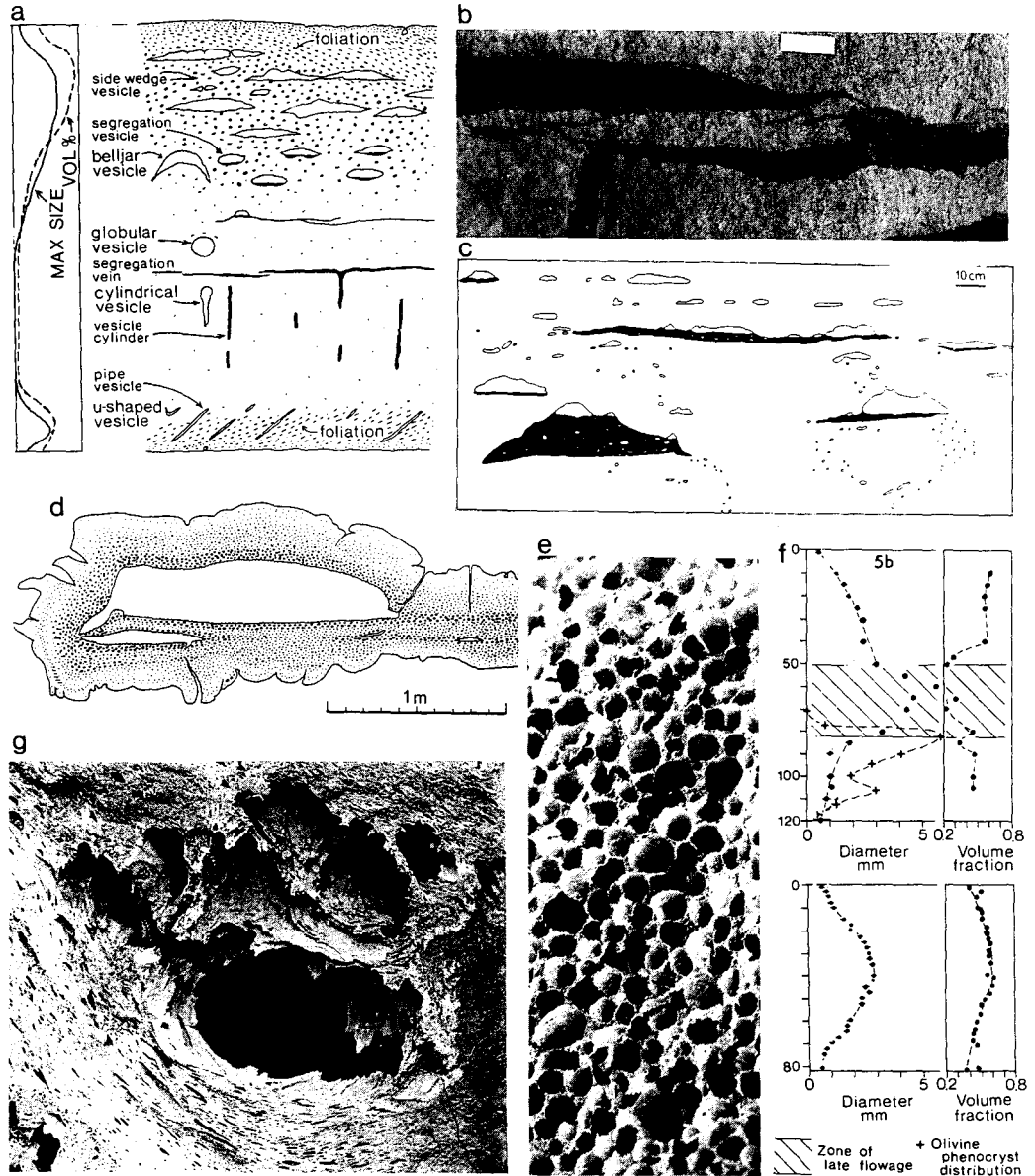
Fig. 8. Vesicularity of basaltic rocks.

(a) Generalized section across a pahoehoe flow unit about 5 m thick of Xitle volcano in Mexico, showing varieties of megavesicles and their zonal distribution (Walker 1993); profiles of vesicle size and abundance left. (b) Segregation vein (dark) in basalt flow, E Iceland; scale is 15 cm long. (c) Segregation vesicles with zeolite-infilled amygdales, Ballintoy, Antrim. (d) Gas blister in a pahoehoe flow unit, Reykjanes, Iceland. (e) Spongy pahoehoe, La Palma, Canary Islands; vesicles up to 5 mm in size. (f) Profiles across two spongy-pahoehoe flow units, Hawaii, showing variations in size and volume fraction of vesicles; depth scale in cm; olivine crystals are strongly concentrated in the lower part of one unit (Walker 1989b). (g) A type of vesicle that grew dynamically in a fast-moving aa flow (Xitle volcano, Mexico).

blisters are distinguished from drained lava tubes because they have bubble-wall texture on roof and floor, and the floor lacks flow structures.

Thick flow units of pipe-vesicle bearing ('P-type') pahoehoe that occur in Hawaii on slopes <4° have fewer and larger vesicles due to longer residence of the lava in tubes (Wilmoth & Walker 1993). Significant bubble loss has occurred. A banding of vesicles often occurs due to growth of the lava by continued injection under a surface crust ('lava-rise' mechanism).

Aa flows have considerably different vesicle distribution patterns because active flowage persists until considerable cooling has occurred. Shearing progressively eliminates vesicles, and distal-type aa tends to be non-vesicular apart from new planar crevices that are caused by shearing.

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