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SUMMARY

The remarkable swarm of inclined basic intrusive sheets found in the deeply dissected part of SE Iceland is believed to constitute part of a widespread layer which, in other parts of the country, occurs below sea level as the 6·35 km s^{-1} crustal layer revealed by seismic refraction studies. A simple mechanism for the generation of the sheet swarm is proposed, based on the contrasted density gradients of crust and uprising magma: basic magma rises to the surface when its density is everywhere less than the bulk density of the rocks it cuts, otherwise it is often diverted laterally to form an intrusive sheet where its density equals that of the country rock. Once initiated, density relations in the crust are such that the swarm is strongly self-perpetuating in nature. It is believed that more than half of the uprising magma in Iceland has been so diverted to form the sheet swarm. One corollary is that the crust in Iceland "filters" the magmas entering it so that only the lighter or those which rise at a high volumetric rate succeed in passing through to the surface. Confluent sheet gabbro intrusions may develop when the frequency of uprise of magma batches is high. Sheet swarm cupolas or perched swarms also occur in central volcanoes, in which the low-density acid volcanic rocks "capture" uprising magmas.

1. The sheet swarm in southeastern Iceland

THE remarkable swarm of inclined basic intrusive sheets which cuts the Tertiary and early Quaternary volcanic pile in SE Iceland is one of the most striking features of the geology of the country. It has received little attention from geologists and scant recognition in the literature, and the following account is based largely on unpublished field observations made by the author from 1962 to 1966, together with observations made by A. E. Annels from 1964 to 1966 (Annels 1967). The swarm is exposed over about 1700 km² (Fig. 1), and intrusive sheets constitute more than 10% of the total rock over one third of that area.

The salient features are as follows:

1. Where most concentrated, sheets make up more than 80% of the total rock. This high concentration is found over 100 km² or more and is therefore not an unusual or isolated occurrence. The concentration of sheets is typically greater

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than that of dykes: sheets commonly constitute more than 20% of the total rock, whereas dykes seldom do.

2. Individual sheets are typically 0·2 to 2·0 m thick, and the swarm contains many tens of thousands of members. The sheets are chilled against the rocks they cut, whether lavas or earlier sheets. Their grain-size is a function of thickness: the thinnest sheets are basaltic, and the thickest gabbroic.

3. The sheets form a parallel set inclined towards the spreading axis in the same direction as, but more steeply than, the country rocks.

4. Traced down-dip (Fig. 2), sheets increase rapidly in number. The swarm increases in concentration downwards (i.e., inversely with height above sea level) from a very low value (less than 1% of the total rock) to a very high one (more than 50%) over a few hundred metres of vertical height.

5. The dip of the lavas and associated volcanic rocks is closely related to the concentration of sheets, as is seen from a comparison of Fig. 3 with Fig. 1, being less than 10° where sheets are scarce or absent and rising commonly to 15°–25° where sheets are abundant. Where sheets are few, they approach parallelism with the stratification of the volcanic rocks, but traced down-dip the dip of the sheets increases more rapidly so that where they are most abundant they dip at 35° to 50°, about 20° to 25° more steeply than the lavas.

6. The sheets are discordant, injected mostly along newly-created fractures which in detail are highly irregular.

7. A large proportion of the sheets are porphyritic and carry abundant bytownite phenocrysts. The proportion is much higher than the 12% of porphyritic basalts found in the Tertiary lava pile of E. Iceland.

8. Most of the sheets are basic, but acid and intermediate members also occur particularly in the country between Hornafjordur and Lon, and tend to be much thicker than the basic ones. Many are composite and possess basic margins.

9. Although the sheets cut all kinds of rocks, basaltic hyaloclastites are particularly conspicuous where the swarm is most concentrated, for example at Hoffellsjokull and the inner part of Kalfafellsdalur. Hyaloclastites appear to attain their maximum thickness of about 1000 m in the relatively inaccessible country north of Dalsheidi, and are attributed to vigorous intraglacial basaltic volcanism in the early Quaternary ice-sheets.

10. The sheet swarm roughly coincides with a widespread zone of green propylitic alteration of the rocks, and the top of the swarm is roughly parallel with the zeolite zones in the lavas above (Walker 1961).

11. Most of the known Icelandic gabbro and granophyre intrusions occur in the area of the sheet swarm.

2. A proposed emplacement mechanism

The geological structure of Iceland indicates that down-sagging along the spreading axis is an important process, responsible for the dip of the lavas towards the axis. The fact that the dip is greatest where the sheet swarm is most concentrated suggests that the injection of the sheet swarm is partly responsible for the down-sagging or tilting. A very simple mechanism is now proposed for the emplacement of the sheet swarm, in which down-sagging is partly the consequence of sheet injection.

This mechanism is based on the fact that the density of basic magmas is higher than that of many volcanic rocks, in particular young basaltic lavas in which the vesicles and the spaces in the auto-brecciated zones are still empty, or hyaloclastites and pillow lavas formed by eruptions in water.

The density of basaltic glass is 2·82 to 2·90 g cm^{-3} at 20°C, falling to 2·59 to 2·69 at magmatic temperatures (Tilley 1922; George 1924; Dane 1941; Bottinga & Weill 1970). From these values, 2·65 for basaltic magma has been widely adopted for basic intrusions (*e.g.* Hess 1960) but 2·73 is taken by Macdonald (1963) for Hawaiian basaltic magma and about 2·70 agrees more with the upward concentration of bytownite phenocrysts (density 2·69 g cm^{-3} at magmatic temperature) found in some intrusions (Blake 1968) and many lavas in Iceland, although the phenocrysts in the lavas may admittedly have been buoyed up by attached gas bubbles.

As basaltic magma rises its density changes (Fig. 4a) due mainly to the onset of vesiculation at level V by the exsolution of dissolved gases. The depth of V depends on the amount of gas but may normally be less than about 1 km. Moore & Schilling (1973) found that undegassed pillows dredged from the Reykjanes Ridge average 0·33% of water and should start to vesiculate at a depth of much less than 1 km (Sigvaldason 1968). Different batches of basaltic magma, of different volatile or phenocryst contents and chemical compositions must give different depth/density profiles, and most of them will plot within the stippled field of Fig. 4a. In particular, a partially crystallised magma (one bearing phenocrysts) should have a higher density than a wholly liquid one. Now consider the density of the crust in Iceland.

The density of young solidified basaltic lavas varies from around 2·9 or 3·0 g cm^{-3} for the dense flow interiors to 2·0 or less for many of the highly vesicular or auto-brecciated layers and for intercalated sediments or ashes. The average density may vary from 2·7 to 2·8 in parts of a lava pile where the flows are thick and massive, to less than 2·5 in parts made of hyaloclastite, pillow lava or the thin and highly vesicular flow units of compound pahoehoe lavas.

The best available picture of the crustal density in Iceland is that from the seismic refraction profiles of Palmason (1971). The average thickness and P velocity of each of the five crustal layers recognised, together with the inferred average density values of the layers, are listed in Table 1, and given as the smoothed profile (curve C in Fig. 4b). Now compare this with magma density profiles.

TABLE 1 *Crustal Layers in Iceland*

Layer	Thickness km	P velocity km s^{-1}	Density g cm^{-3}
0	0·5	2·8	2·1–2·5
1	1·0	4·2	2·6
2	2·15	5·1	2·65
3	6	6·5	2·9
4 (mantle)	—	7·2	3·1

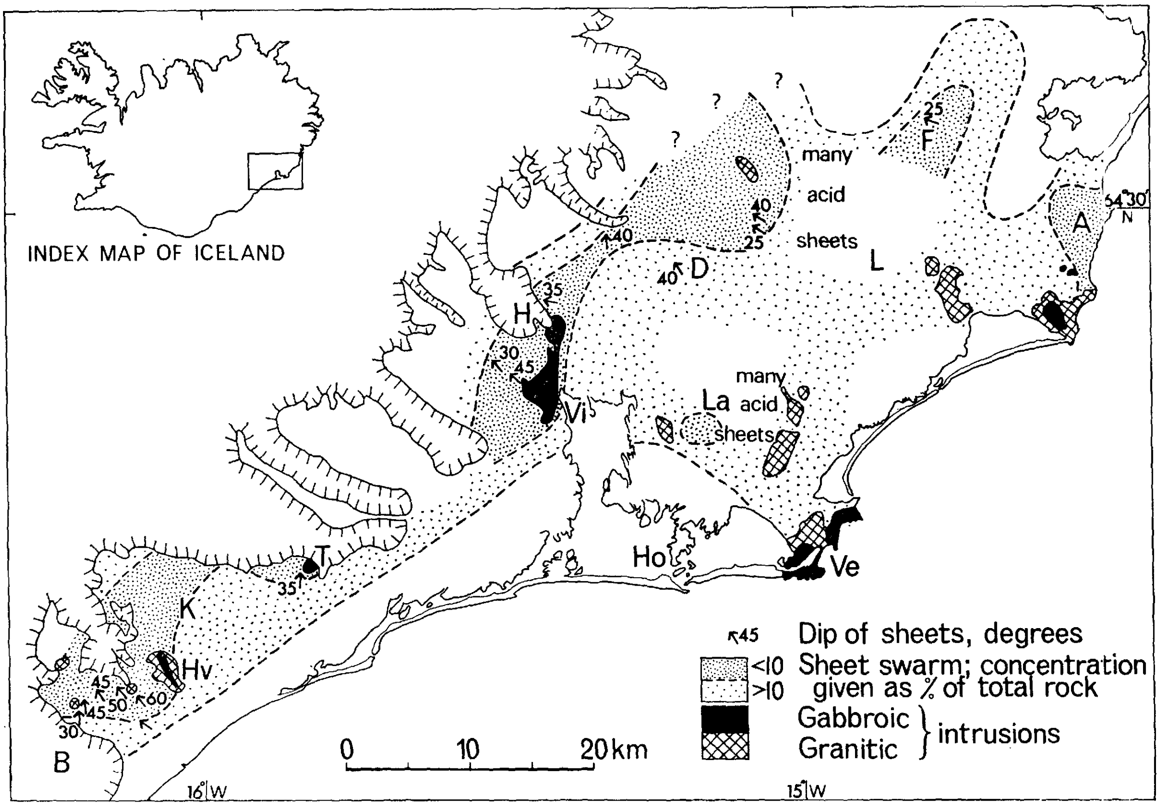

INDEX MAP OF ICELAND
many
acid
sheets
many
acid
sheets
A
F
D
L
H
Vi
La
T
Ho
Ve
K
Hv
B
64°30' N
16° W
15° W
0
10
20 km
Dip of sheets, degrees
<10
>10
Sheet swarm; concentration given as % of total rock
Gabbroic
Granitic
intrusions

FIG. 1.

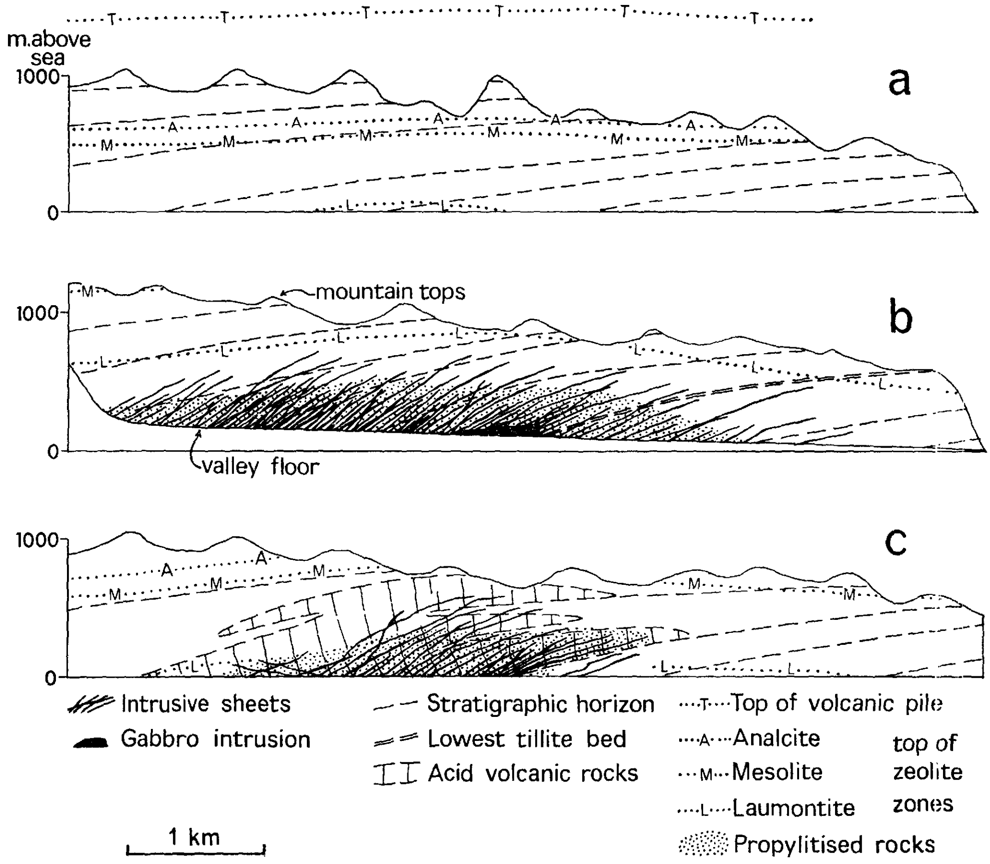

m. above sea
1000
0
a
b
c
mountain tops
valley floor
Intrusive sheets
Gabbro intrusion
Stratigraphic horizon
Lowest tillite bed
Acid volcanic rocks
Top of volcanic pile
Analcite
Mesolite
Laumontite
top of zeolite zones
Propylitised rocks
1 km

FIG. 2.

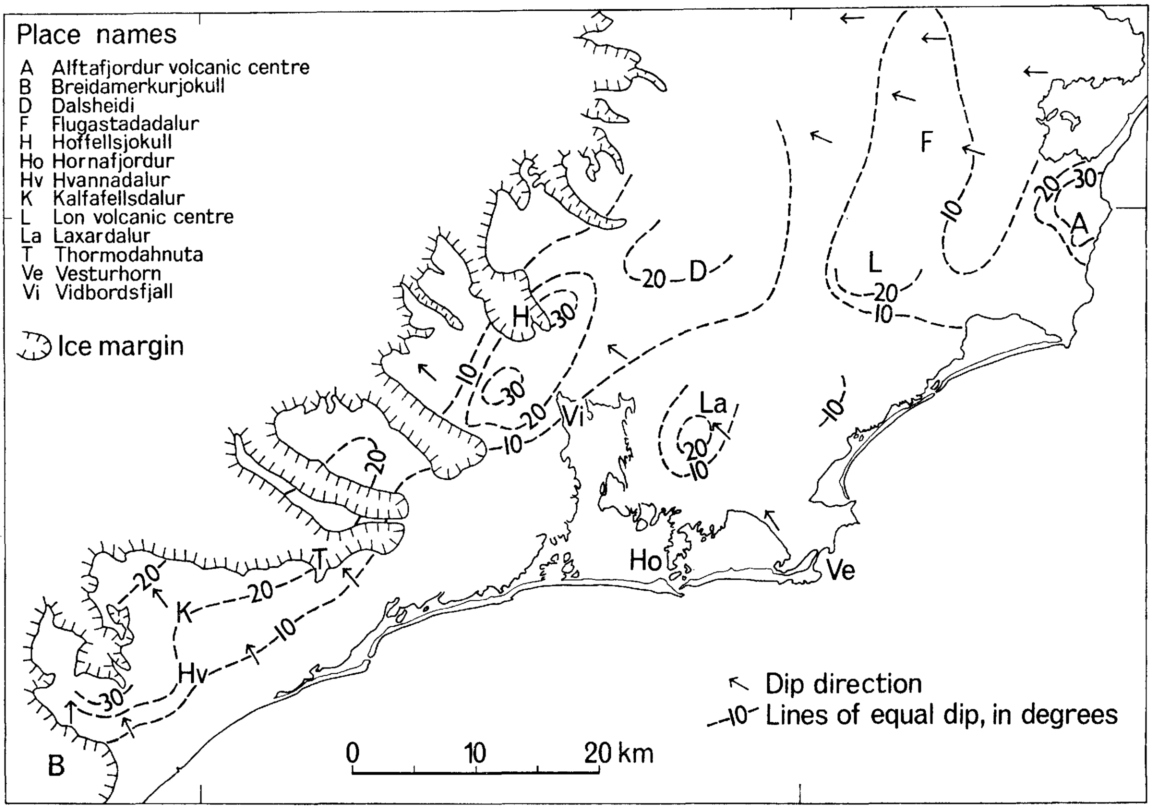


FIG. 3. The variation in dip of the lavas at or near sea level for the same area as Fig. 1.

The density profiles of two typical basaltic magmas are superimposed on the crustal profile of Fig. 4b. One, B, intersects the crustal curve at a depth of about 3 km and the other, A, fails to intersect it. There is of course much uncertainty in the form of each profile and hence as to whether, or where, intersection will occur; the point is that from data currently available the density curves for basaltic magmas and the upper crust are so close that the curve for some basaltic magmas is likely to intersect the latter, and at a depth of several kilometres. Now consider the significance of this.

FIG. 1. Distribution of the intrusive sheet swarm in south-eastern Iceland. The concentration values apply at or near sea level. The inset map shows the location of the area; individual localities are lettered, the explanation of which is in Fig. 3. Dense stipple indicates where sheet concentration exceeds 10%.

FIG. 2. Schematic sections typical of:

a. E. Iceland, where the sheet swarm is not exposed at the present erosion level
b. A valley in the sheet swarm of SE Iceland showing the main feature of the swarm
c. A central volcano, showing a possibly "perched" swarm of intrusive sheets localised in the acid volcanic rocks there.

Along a crustal spreading axis batches of basaltic magma rise periodically through the crust under hydrostatic pressure along fissures, and two possibilities then arise. In one, the magma has a density which is everywhere lower than that of the volcanic crust, A, Fig. 4b. It rises to the surface and a volcanic eruption ensues. In the other, B, Fig. 4b, the magma has a density curve which crosses that of the volcanic rocks. When it reaches the level at which its density equals that of the crust it may cease to rise, but may instead move laterally along an equi-density surface as an intrusive sheet.

The idea that a magma may float rocks less dense than itself or more correctly flow underneath them is not new, and has been discussed in connection with the origin of laccoliths and sills by many including Gilbert (1877) and Bradley (1965). The process was watched by the author during the eruption in Iceland in February 1973 when half of the new scoria cone of Eldfell moved laterally, evidently on a sole of basaltic magma which had flowed beneath the cone.

The variation in lithostatic pressure, P_L, through the crust is plotted on Fig. 5a, taking Palmason's average values for the thickness and density of the crustal layers. The hydrostatic pressure, P_H, for magmas of densities 2·5, 2·6, 2·7 and 2·8 g cm^{-3} is also plotted, the simplifying assumption being made that the crust rests on a yielding substratum so that the hydrostatic pressure in magma at the base of the crust is equal to the lithostatic load there. This may not be an unreasonable assumption since recent work (Hermance & Grillot 1974) suggests that magmatic temperatures occur at a depth of 10 to 15 km below Iceland.

Figure 5b plots the excess hydrostatic pressure, P_{He}, the amount by which the hydrostatic pressure exceeds the lithostatic, as a function of depth for the same four magmas. For each, P_{He} reaches a maximum at the level where the density of the magma equals that of the crust. It is postulated that a sheet is intruded at this level, P_{He} being the force which drives the sheet forward.

The author believes that the reason why the magma does not reach the surface is that the highly permeable and (below a generally high water table) waterlogged surface volcanics constitute a thermal barrier which many batches of magma are

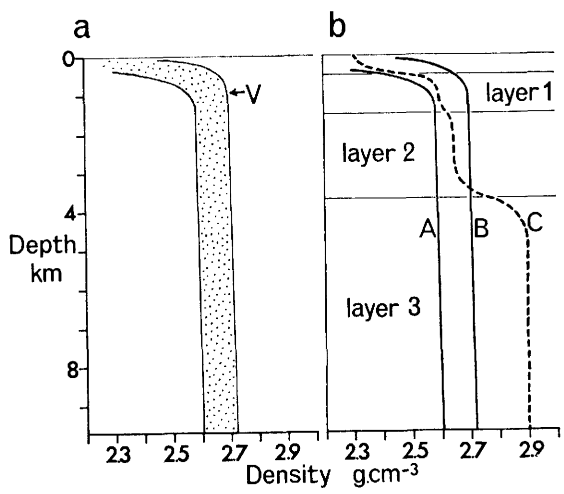


FIG. 4.

a) The density variation of common basaltic magmas with depth. Some magmas rich in mafic phenocrysts, uncommon in Iceland, may lie to the right of the stippled field. V is the level at which vesiculation begins.

b) A and B, density profiles for basaltic magmas, superimposed upon that (C) for the crust in Iceland (after Palmason).

unable to penetrate; instead they congeal in the upper part of the fissure. There is much circumstantial evidence for this barrier; for example solfataras were active for two weeks on Askja before the 1961 eruption began (Thorarinsson & Sigvaldason 1962); and many dolerite dykes in eastern Iceland show an upward chilling to sideromelane multiple intrusions.

Intrusive sheets in SE Iceland transgress upwards across the lavas in a direction away from the spreading axis, more steeply near the axis than farther away, and if they were intruded along equi-density surfaces these surfaces must have been similarly inclined. It is not easy to account for this. Three factors that could contribute are illustrated in Fig. 6. The non-uniform distribution of dykes is thought to be the main factor; the effect of temperature, and density changes due to the infilling of voids by zeolites, are thought to be small.

There is another significant relationship: as the magma in a sheet rises up-slope its density decreases, particularly if vesicles start to form, and the sheet may then cross equi-density surfaces and move towards crustal rocks of lower density so that it becomes inclined more steeply than these surfaces (Fig. 6e). Vesicles do occur in a proportion of the sheets, although generally in small amounts.

The initiation of a sheet swarm is not easy to account for, but once it has been initiated it must so change the density distribution in the upper crust that later sheets are almost inevitably injected along or near the top of the existing swarm: the swarm is strongly self-perpetuating in nature.

Acid sheets also occur and are locally abundant. The density of acid magma, 2·1 to 2·3 g cm^{-3}, is so low that the magma should everywhere in Iceland be capable of reaching the surface. The problem is then one of finding a mechanism, firstly by which it can rise as an inclined sheet, and secondly to restrain it from reaching the surface (necessary in any intracrustal intrusion). Regarding the first, acid magma has too low a density to form an inclined sheet in the same way as basic, but it can rise up the thermally favourable pathway in the middle of a still-hot basic sheet (*cf.* Gibson & Walker 1963). Many of the acid sheets are,

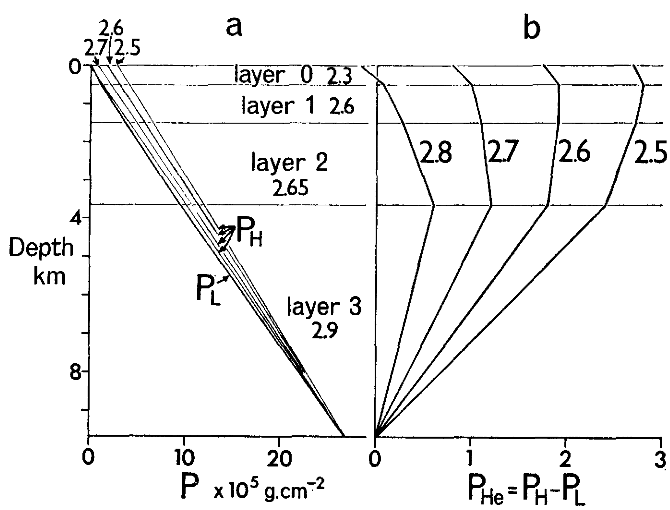


FIG. 5.

a) Profile through the crust in Iceland showing, P_L the lithostatic pressure and P_H, the hydrostatic pressure for magmas of densities 2·5, 2·6, 2·7 and 2·8 g cm^{-3}.

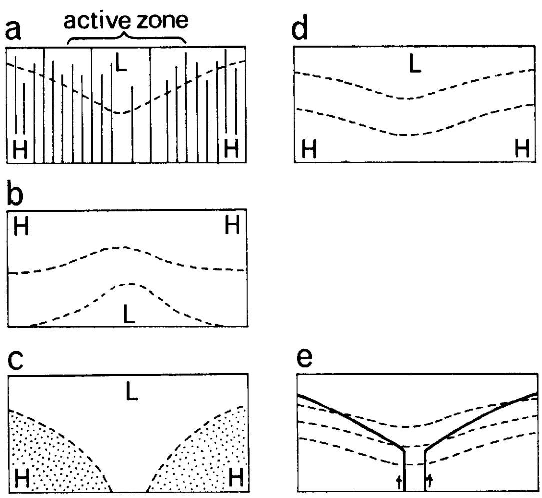
b) The excess hydrostatic pressure for the same four magmas.

in fact, composite and possess basic margins. Regarding the second requirement, the rapid viscosity increase when the acid magma reaches the end of a hot basic pathway and enters cold rocks will normally restrain it from travelling much farther unless the magma body is a large one.

The above emplacement mechanism for the sheet swarm differs from that proposed by Piper & Gibson (1972) in which the stress field is thought to be the prime control. One test of the two mechanisms is a petrological one: a large proportion of the sheets are porphyritic, which is taken to imply that their magma had a relatively high density. Another test is environmental: a large proportion of the sheets are emplaced in relatively low-density rocks such as hyaloclastites. Neither relationship can be explained by the stress control mechanism; both can be explained by the density control one.

3. Uplift or subsidence by sheet intrusion

It has commonly been tacitly assumed that the space occupied by a basic sill or intrusive sheet is made available wholly by an uplifting of the superincumbent rocks, and magmatic pressures have therefore been invoked sufficient to lift the overburden. In fact it seems more likely that most of the space is made available by a subsidence of the subjacent rocks. The proportion made available by each is determined by the density relationships of the intrusion and the supposed yielding sub-crustal substratum. Fig. 7 shows for different densities of substratum the net uplift of the surface, Sm, produced by the intrusion of a sheet of magma of vertical thickness 112 cm and density 2·7 g cm^{-3} and also that resulting, Sc, after the same sheet has solidified to a crystalline dolerite 100 cm thick of density 3·05 g cm^{-3}. If the substratum has a density of 3·1, which seems a realistic value



a. Non-even distribution of dykes across the width of the active zone
b. Density differences due to the temperature of the rocks
c. Density increase due to the infilling of voids by zeolites in the stippled parts
d. Shows the position of equi-density surfaces, the resultant of a to c
e. Illustrates how sheets might cross equi-density surfaces if the density of the magma decreases as it rises, for example because of vesiculation.

FIG. 6. a to c. Schematic sections across the active zone in Iceland to show the factors which might contribute to produce non-horizontal equi-density surfaces. In each, H and L are the areas of highest and lowest density resulting from the effect specified.

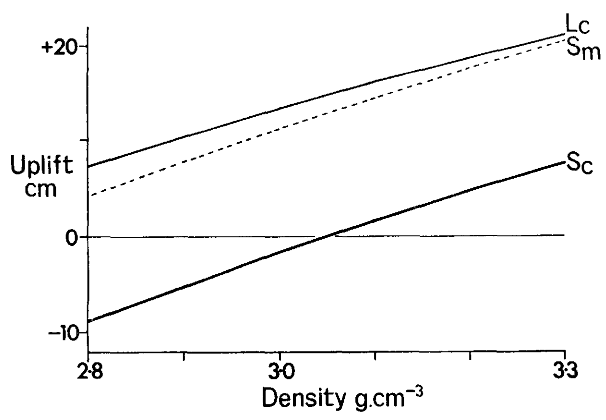
for Iceland, the ground above the sheet is raised by 14 cm, but this uplift is reduced to only 2 cm when the sheet has crystallised. In addition a local though presumably temporary subsidence might result by the withdrawal of magma from below.

Fig. 7 also shows the net raising of the land surface, Lc, which would result when a flow of solidified lava of thickness 100 cm and density 2·6 g cm^{-3} is erupted. For a substratum of density 3·1, the floor on which the lava rests would subside by 84 cm and the land surface would stand 16 cm higher than before. This subsidence can account for at least part of the dip of the lavas found almost everywhere in Iceland; the intrusion of the sheet swarm may account for the remainder.

4. The regional nature of the sheet swarm

It is thought that the sheet swarm is exposed in SE Iceland because of the deep level of erosion there; that it constitutes a widespread and integral part of the structure of Iceland and is not merely a local phenomenon; and that over most of the area of Iceland it is not seen because it lies below sea level.

There are three reasons for believing that the area where the swarm is exposed is the most deeply dissected in Iceland. Firstly it includes some of the deepest valleys, with a relief of as much as 1200 m, while in most places elsewhere the relief is not more than 800 or 1000 m. Secondly, the zeolite zones are highest there. The top of the laumontite zone (normally 1400 m below the top of the lava pile (Walker 1961) and below sea level in most of Iceland) rises to more than 500 m above sea level. Epidote, normally a still deeper mineral, is found where the sheet swarm is highest at 1100 m on the mountain summits just north of Breidamerkurjokull. The possibility admittedly exists that the thermal gradient may have been unusually high during the period of zeolitisation in this part of Iceland, but even so the land surface at that time can hardly have been less than 2000 m above present sea level. Thirdly, most known Icelandic intrusions of gabbro and granophyre occur in this part of Iceland. They are very young; some are only



Sm, the intrusion of a sheet of basic magma 112 cm in vertical thickness and density 2·7 g cm^{-3};
Sc, the same intrusion solidified to a dolerite intrusion 100 cm thick and density 3·05;
Lc, net raising of the land surface by the eruption of a basalt flow 100 cm thick of density 2·6.

FIG. 7. Diagram showing, for different densities of yielding substratum, the net uplift of the land surface resulting from:

2 m.y. old (Gale *et al.* 1966), but the rate of erosion is known to be exceptionally high. Thus the present rate of down-cutting by the Hoffellsjokull is at least 5·5 mm y^{-1} averaged over the whole area of the glacier (Thorarinsson 1939), and probably several times that for the part which occupies a deep valley.

The exposure of the swarm in SE Iceland may not be due entirely to deep erosion there: the swarm may itself have reached a higher than normal level in the crust because of the unusual development of hyaloclastites there. When formed, the hyaloclastites were rich in basaltic glass and had a high percentage of inter-grain voids, to give a bulk density much lower than 2·5 g cm^{-3}. The voids were soon infilled and some glass crystallised, but still the bulk density was relatively low. The thick hyaloclastites would thus have been an ideal site density-wise for the emplacement of sheets.

Some sheets cutting hyaloclastites at Hoffellsjokull are brecciated. It is not certain whether they nearly reached the surface and were the source of some of the hyaloclastites, or merely penetrated and were chilled and broken against loose and waterlogged hyaloclastic breccias.

5. Sheet swarms in central volcanoes

A feature of volcanism in Iceland is the presence, within the flood basalt pile, of central volcanoes wherein are concentrated acid and intermediate volcanic rocks. These volcanoes were first recognised in E. Iceland and nearly 30 dissected Tertiary–Quaternary examples are now known (Sigurdsson 1967) of which nine have been described in detail.

The core region in each volcano is cut by inclined intrusive sheets (mostly basic, although intermediate and acid members also occur) forming swarms remarkably similar with one another and with the regional swarm. They are generally 1 m or less thick, and tend to be inclined towards the centre of the volcano.

A difficulty is that at any one exposure dykes and other sheet-like intrusions typically trend and dip in many different directions, and even a single intrusion may vary greatly over the space of a few tens of metres. It is not easy objectively to distinguish a set of inwardly inclined sheets from other intrusions without a rigorous statistical treatment of strike and dip data. Moreover the dip of the country rocks also varies greatly, and it is not always possible either to ascertain whether it is an original depositional slope rather than due to subsequent tilting, or whether a particular intrusion was emplaced before, during, or after an episode of tilting.

Some sheets in the Reydarfjordur volcano are sill-like (Walker 1959) and some cross-cutting (Gibson *et al.* 1966). Inclined sheets, common at Breiddalur (Walker 1963), Thingmuli (Carmichael 1964) and Alftafjordur (Blake 1970) were not called cone sheets because their centrally-inclined tendency is weak. Swarms of what are described as cone sheets occur at Kroksfjordur, Stardalur and Vididalur–Vatnsdalur (Annells 1969). In the first volcano (Hald *et al.* 1971) they are distributed around 120° of arc and for cone sheets have a low dip (17° to 41° shown on the map, *loc. cit.*). Where most concentrated they make up 10 to 15%

of the total rock. In the second volcano (Fridleifsson & Kristjansson 1972) they are known to be distributed around 130° of arc and dip inwards at about 35° but occur within a caldera cutting rocks that have an average inward dip of about 25°. In both volcanoes the sheets are mostly concentrated in the quadrant facing the spreading axis.

These sheets have perhaps closer affinities with the regional swarm in Iceland than with the type cone sheet swarms of the British Tertiary. "Centrally inclined sheet" is perhaps preferable to "cone sheet" because no one sheet is known to extend around more than a few degrees of arc (this however is true of the British cone sheets too) and also because "cone sheet" has come to have definite implications, probably not intended by the originators of the term, regarding their mechanism of formation. In Iceland it is also convenient to refer to the sheets in the volcanoes as "centrally inclined," and those in the regional swarm as "axially inclined."

The more complete cone sheets swarms at Setberg (Sigurdsson 1966), being in a tectonic setting different from the other Icelandic examples and more comparable with that of the British Tertiary, are not discussed here. Now consider the origin of the swarms in the volcanoes.

Each volcano has a concentration of low density acid rocks in it, typically constituting 25 to 50% of its bulk (Walker 1964) and including lavas (density about 2·5 g cm^{-3}) and pyroclastic rocks (density much less than 2·5) in comparable amounts. Each volcano also has a down-sagged core region of high dips only occasionally (as at Stardalur) fault-bounded. The sheets and associated dykes may be the prime cause of the down-sagging. The intrusions which are exposed are generally too few to account for the observed subsidence, but the concentration of sheets increases rapidly downwards and it is inferred that it could reach a high value a few hundred metres lower down.

It is now postulated that the sheet swarms in the volcanoes were formed by the same mechanism as in the regional swarm, and that their localisation was due directly to the low density of the acid volcanic rocks in and near the centres. The centrally inclined tendency is partly determined by the spatial distribution of acid rocks, and partly by central down-sagging.

The topographic level at which a sheet swarm occurs in a volcano is typically one at which sheets do not occur in the flood basalts on either side. The question then arises whether the swarm occupies a cupola-like volume projecting above the

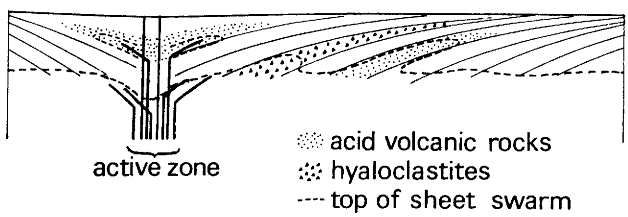


FIG. 8. Schematic section across the crustal spreading axis in Iceland showing where the regional swarm of intrusive sheets and the swarms in the central volcanoes may originate. Some of the sheets in the latter may be emplaced after the volcano has left the spreading axis.

general level of the regional swarm, or is "perched" (like a perched water table) and completely disconnected from the regional swarm below, or lies between these two extremes and occupies a mushroom-shaped volume. Some evidence that it is mushroom-shaped or perched comes from Thingmuli and Breiddalur where sheet-bearing parts of the central volcano overlie flood basalts containing few or no sheets (Fig. 2c).

The sheets in the volcanoes differ from those in the regional swarm in that few are porphyritic. The explanation may be that a wider spectrum of magmas can be trapped by the low density acid rocks, and porphyritic examples are relatively scarce because they have been "diluted" by a relative abundance of other types. Alternatively and more likely it is because the porphyritic magmas, being relatively heavy, have formed sheets lower down and have not entered the central volcanoes.

6. The identity of Crustal Layer 3 in Iceland

Of the four main crustal layers, layers 0, 1 and 2 of Palmason are confidently interpreted as basalt lavas and associated rocks, but the identity of layer 3 still poses a problem. The thermal gradient from boreholes in southern and western Iceland suggests that the temperature at the top of 3 is 350 to 400°C, suggesting that 3 is a metamorphic facies of volcanic rock, probably amphibolite (Palmason 1971). This is consistent with the relative horizontality of the 2–3 boundary compared with the dips of the surface Tertiary rocks. The shallowest depths are found in volcanic centres, and the metamorphism model is consistent with this also since the geoisotherms are likely to have risen to shallower depths below these centres when they were active.

The alternative is now proposed that layer 3 consists of a very intense regional swarm of intrusive sheets of which that exposed at the surface in southeastern Iceland is a part. Such a swarm could give the appropriate seismic velocity, and each volcanic centre moreover has a sheet swarm at shallower depth; in short, the identity of layer 3 with a sheet swarm has all the merits of the metamorphic zone model. It could be tested by measurements of seismic velocity made within the sheet swarm in SE Iceland: existing seismic profiles lie entirely outside the swarm or in parts of it where the concentration of sheets is very low.

This swarm is remarkably similar to the sheeted complex in the Troodos Mountains of Cyprus, a complex which constitutes a distinct layer in what is widely acknowledged to be a portion of ancient oceanic crust (Gass 1968; Gass & Masson Smith 1963). Seismic velocities measured in the near-surface part of the sheeted complex (Matthews *et al.* 1971) are relatively low—the average in it and the associated gabbro is 5·2 km s^{-1}—but this may be due to the presence of open cracks at or near the surface. The sheets in Cyprus have a steeper dip than those in Iceland, but layer 3 in Iceland and the sheeted complex in Cyprus are equated with one another.

7. The effect of the rate of uprise of magma

The fact that fissure eruptions take place periodically as short events separated by long time intervals and that individual intrusive sheets or dykes are chilled against the rocks they cut indicates that basaltic magma rises periodically into the upper part of the crust in Iceland as discrete batches. Several relationships can then be explained as depending upon the volumetric rate of uprise of this magma. Three kinds of rate are envisaged: the frequency (F) of uprise of magma batches, the total volume (V) of each, and the rate of delivery (D) of the batch.

Little is known of the magnitudes of these three except from recent surface eruptions, for which F, expressed as the number of eruptions per thousand years per 20 km length of active zone, ranges from less than 1 to more than 2; V averages about 0·1 km^3 but can rise to as much as 15 km^3 in individual examples; and D, averaged for the whole duration of an eruption, probably lies mostly in the range 5 to 500 m^3 s^{-1}. For the intrusive sheets, if the non-eruptive fissures (gjas) common in the active zones indicate batches of magma which failed to reach the surface, F might have a value several times greater than for surface extrusions, and from the thinness of the sheets V might well be less than the 0·1 km^3 for extrusions.

One possible relationship stemming from the rate of uprise is that a high D value may favour surface eruptions. When the pathway to the surface clogs, as is likely to happen when D is small, then conditions favour the formation of a sheet instead.

A second relationship is suggested by the gabbro intrusions such as are found rather intimately associated with the sheet swarm. The distribution of the main gabbros is shown in Fig. 1; in addition some of the larger intrusive sheets (those thicker than about 10 m) which are not shown on the map have a gabbroic centre.

For a basaltic or doleritic sheet to form, the rocks into which it is intruded should be relatively cold—F should be small—and V should presumably also be small if the sheet is to be a thin one. If the value of F then rises to a high value so that an incoming batch of magma arrives before the preceding sheet has fully solidified, then the new magma is likely to be preferentially intruded into this sheet since less energy is required to follow this pathway than to create a new one. When this is repeated several times in quick succession the resulting thick multiple sheet will start to take on the characters of a single thick intrusion, since the time for complete solidification is proportional to the square of the thickness (Jaeger 1958). Injection repeated many times produces a gabbro intrusion which is likely to show a form of layering due to multiple injection and may also show layering due to crystal accumulation. It may locally have internal chill zones of doleritic grain-size. One can speculate that, once started, the growth of a gabbro intrusion is a self perpetuating process which will be brought to an end only by F falling to a low value again.

Some gabbro intrusions in SE Iceland show features indicative of multiple injection (Newman 1967). They contain screens of country rock, and sheets of gabbro and dolerite come off from them. Like most of the thinner sheets they tend to be rich in bytownite crystals.

The effect of variations in V is uncertain, but presumably a large V value results in a sheet both extending farther laterally and having a greater thickness. Once a gabbro intrusion has commenced to grow however, an increase in V will merely hasten its growth.

8. Conclusions and corollaries

A major swarm of intrusive sheets is recorded from SE Iceland and a simple mechanism is proposed to account for it and for crustal layer 3 with which it is equated. The mechanism can account for many apparently unrelated phenomena which tends to support its broad validity:

(1) The gaping dilation fissures or *gjas* which are common in the active zones of Iceland may be eruptive fissures in which the magma failed to reach the surface (Bodvarsson & Walker 1964; Walker 1965), and can be explained as the surface expression of the diversion of uprising batches of magma to form intrusive sheets.
(2) The surface deformations measured by Tryggvason (1968) from the active zone at Thingvellir, where no surface eruptions have occurred for more than 1000 years, and at Reykjanes (Tryggvason 1970) where few have occurred could be accounted for by the injection and solidification of intrusive sheets.
(3) The excess heat flow in Iceland over the global average level can be accounted for by the emplacement of intrusions having a total volume about four times greater than that of surface extrusives (Bodvarsson & Walker 1964), only some of which could be due to feeder dykes. The heat transfer from intrusive sheets to meteoric waters will take place more readily than from large intrusions, and the frequent injection of sheets is capable of maintaining the excess heat flow in the high temperature thermal areas even where no surface volcanism has occurred over the past 10^3 to 10^4 years.
(4) A high-concentration sheet swarm appears fully capable of satisfying the seismic velocity, density and depth of layer 3, and the downward increase in concentration is thought to be sufficiently rapid to produce the 2–3 boundary.
(5) The relatively high level sheet swarms in the central volcanoes and their wide compositional range can be explained by the concentration of low-density acid volcanic rocks there.
(6) The zeolite zones in the lavas are roughly parallel to the top of the sheet swarm, and some at least of the thermal energy required to initiate or sustain zeolitisation may have come from the sheets.
(7) The prevalent dip of the volcanic rocks in Iceland is attributed to subsidence accompanying the eruption of lavas and the intrusion of sheets.

The following are some corollaries which could be used to test the validity of the mechanism:

(1) Surface extrusions sampled for the purpose of petrological or geochemical studies are not representative of the magmas which rose into the crust, since those magmas which had a relatively high density (by virtue of their ferromagnesian-rich or gas-poor character or the abundance in them of phenocrysts), or which rose at a low volumetric rate, have been diverted before they reached the surface to form intrusive sheets instead.
(2) Lateral petrographic or geochemical variations along the length of a spreading axis could reflect lateral variations in the density of the upper part

of the crust insofar as they control the spectrum of magmas which are able to pass through to the surface.

(3) During the latter part of, and after, a long period of glaciation the vigour of surface extrusive activity may be greatly reduced because of the resulting thick piles of relatively low-density hyaloclastites into which most of the batches of uprising magmas are diverted.

(4) A lack of surface volcanism in part of an active zone, as at Reykjanes and Thingvellir, cannot be taken necessarily to imply a lack of magmatic activity: it may merely mean that the uprising magma is being diverted to form intrusive sheets. Paradoxically, a complete lack of surface volcanism could even be due to an increase in the vigour of magmatic activity, producing gabbro intrusions instead. Perhaps a better index to the amount of magmatic activity is the heat flow and rate of surface deformation in the active zone.

(5) A lack of basaltic volcanism on a central volcano could be due to the acid rocks "capturing" rising basic magmas. Indeed a buried central volcano might even be detected by the absence of basaltic activity above it. The Quaternary volcano of Torfajokull (Saemundsson 1972) is an example of one with few basaltic eruptions but appreciable acid volcanism and thermal activity, and the positive Bouguer anomaly indicates the presence of dense rocks below.

(6) The more olivine-rich lavas in Iceland tend to form lava shields rather than extensive lava flows of flood basalt facies, attributed (Walker 1971a) to their generally lower effusion rate. This lower rate could be related to the higher density of the olivine basalt magmas and the corresponding lower hydrostatic pressure at which they reach the surface.

(7) The restricted composition range of ocean floor volcanic rocks could be partly due to the relatively low density of the pillow lavas and associated submarine volcanic rocks into which many batches of basaltic magmas are diverted to form sheets. This control is held to be only partly responsible, since at the other end of the scale the low density magmas may be unable to pass through because of their viscosity (Walker 1971b).

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DISCUSSION OF PREVIOUS TWO PAPERS

Dr M. J. Le Bas offered his congratulations to Dr Walker on his lucid presentation of the role of density during magmatic ascent. However, density effects alone cannot explain the majority of continental basalts which have been extruded through less dense materials. That density control can be important is shown at Carlingford and Skye, where layered gabbros intrude as sheets above sialic crust and at the base of the Tertiary lava pile: Bailey's mechanism for ring dyke formation is almost impossible here. Dr Walker likened the inclined sheets of Iceland to cone sheets, but the former are planar and dip to the rift-zone which gives a different stress field from that of the conical sheets in the Hebrides. It seems that cone sheets raise their central mass of country rock as a result of vertical forces, and that this mechanism is well exemplified by high-level carbonatite complexes. How did he reconcile the fact that his model appears to depend on acid preceding basic magma, contrary to the relations at most Hebridean centres? Had he considered the mechanical strength of the crust through which magma penetrates: the nature and extent of joints and fractures in strata along which magma can find weaknesses? Finally, did he distinguish between the different effects, apart from viscosity, of 'wet' and 'dry' magmas, which can either soften the envelope of an intrusive boss followed by deformation and folding, or give brittle-type fracture of the country rocks permitting sheet intrusion?

Dr R. R. Skelhorn: It was very difficult to believe that such a process would be of importance at depths (5–7 km) where the British Tertiary cone sheet complexes originated. The author had not considered the stress distribution which would be imposed on the country rocks by the diapir. Was this because Dr Walker believed that such stresses, and the consequent fracturing, were unimportant? The speaker believes that cone sheet fractures are produced in a rigid crust where the stress is applied in a short time due to the rapid rise of a cylindrical magma chamber. The distribution of the individual cone sheets within the Mull complexes is remarkably similar to fracture patterns produced in experimental studies, e.g. where a glass plate is subjected to an impact which gives a short but intense pulse of pressure (J. E. Field 1964, *Times Sci. Rev.* **51**, 5–9).