NORDIC VOLCANOLOGICAL INSTITUTE 8502
UNIVERSITY OF ICELAND

STRUCTURE, DEPOSITIONAL ENVIRONMENT AND MODE OF EMPLACEMENT OF BASALTIC HYALOCLASTITES AND RELATED LAVAS AND SEDIMENTARY ROCKS: PLIO-PLEISTOCENE OF THE EASTERN VOLCANIC RIFT ZONE, SOUTHERN ICELAND

by

Steffen G. Bergh

Reykjavik 1985

NORDIC VOLCANOLOGICAL INSTITUTE 8502
UNIVERSITY OF ICELAND

STRUCTURE, DEPOSITIONAL ENVIRONMENT AND MODE OF EMPLACEMENT OF BASALTIC HYALOCLASTITES AND RELATED LAVAS AND SEDIMENTARY ROCKS: PLIO-PLEISTOCENE OF THE EASTERN VOLCANIC RIFT ZONE, SOUTHERN ICELAND

by

Steffen G. Bergh

Reykjavik 1985

CONTENTS

1. ABSTRACT

Plio-Pleistocene strata of the Sida and Fljotshverfi districts in southern Iceland consist of a 600 m thick, subglacial/subaqueous volcanogenic complex (Younger Series) dominated by voluminous (up to 30 cubic km) and extensive hyaloclastite sheets and interbedded sedimentary diamictite units, that overlies, unconformably, a succession of flatlying, terrigenous lavas and sedimentary rocks (Older Series). The Older Series, of probable late Tertiary age, was exposed to deep glacial erosion prior to the deposition of the Younger Series.

Each hyaloclastite unit show a regular structural sequence with basalt and/or pillow lava dominated facies at the base, and various hyaloclastite breccias and layered-and large scale crossbedded hyaloclastite tuff at the top. The same change appears in proximal-distal relationships. The breccia contains many large aligned and complex folded basalt bodies and lobes, attesting for emplacement by considerable flowage.

The hyaloclastite flow units were produced by repeted voluminous extrusions of basaltic lava from subglacial (?) fissure vents, located near the present day Laki-and Grimsvotn fissure system of the axial rift zone. Fragmentation of the extruded lava portion by quenching as the overlying ice (?) started to melt, and avalanche of the hyaloclastite and associated basalts due to gravity collaps, followed the initial eruption stage. The composite mass moved like a high-concentration, partly fluidized pyroclastic flow that infilled topographic depressions, and was gradually more and more decomposed during flowage. The final emplacement of the flow occured in a fully subaqueous environment, by the transportation into drowned valleys along the southern coast of Iceland, or alternatively, in very large dammed lakes. This model is confirmed by sediment-petrographic comparisions with the interbedded diamictites, which are poorly sorted mud-pebble-clast rocks with abundant stratification, forset cross beds and paleotopographically controlled thicknesses. The sedimentary material was probably supplied from inland fluvioglacial rivers and resedimented in valley-and coastal shelf deltas, locally in the form of subaqueous sediment gravity flow deposits. Only minor parts are deposited directly from grounded or terminal ice.

2. INTRODUCTION

This investigation deals with the stratigraphy, internal structural organization, depositional environment and mode of emplacement of Plio-Pleistocene hyaloclastites, lava flows and related sedimentary rocks in southern Iceland (Fig. 1). The geochemistry and compositional variation of the volcanic rocks are treated in a separate paper.

Hyaloclastites of the Icelandic Palagonite System are in general considered to be the result of subglacial volcanism (Noe-Nygaard 1940, van Bemmelen and Rutten 1955, Kjartansson 1943, 1959, Sigvaldason 1968, Jones 1970). The deposits are fragmented hyaloclastite breccias, sideromelane tuffs and some intercalated basaltic pillow lavas. The origin and mode of emplacement of the various types of hyaloclastites are, however, not yet fully understood. The general theory is that they were formed either by explosive fragmentation and/or "in situ" granulation of basaltic lavas, due to rapid chilling, in contact with water/ice. The commonest morphologic products of subglacial volcanism, from the Upper Pleistocene/Recent period in Iceland, are "table mountains"(Pjetursson 1900, Peacock 1926, Jones 1970) and "narrow serrated ridges" (van Bemmelen and Rutten 1955). These landforms probably originated by "in situ" build up of fragmental material above a subglacial central crater vent (Kjartansson 1966, Jones 1970), or a fissure (van Bemmelen and Rutten 1955, Saemundsson 1967).

The present work considers a third possible depositional mechanism of hyaloclastite rocks, namely that by continued brecciation of lava during flowage in water, and where the fragmented eruptive masses are spread out as huge hyaloclastite sheets to cover large areas instead of producing narrow emerged ridges. This mode of formation has received relatively little attention in the past, but it has been previously suggested as an importent process of breccia formation (Einarsson 1946, 1962, Walker and Blake 1966). This work also questions the general inferred subglacial origin of hyaloclastite, mainly on basis of depositional reconstructions of intervening sedimentary rocks.

Formation of basaltic hyaloclastites by flowage has been reported from ocean floors (De Rosen-Spence et al. 1980) and seamounts (Schmincke 1982), but few published papers are conserned with subglacial/shallow subaqueous environments. Fuller (1931) described lava flows of the Columbia River Basalts that entered a lake and were granulated by interraction with water during flowage. Similarly, littoral hyaloclastite deltas may form by lavas descending into the sea, as reported by Furnes and Fridleifsson (1974) and Furnes and Sturt (1976). Walker and Blake (1966) described a huge hyaloclastite mass that may have originated by break up of a lava flow that drained down along a valley beneath a valley glacier, at Dalsheidi, in eastern Iceland. Preusser (1976) noted some extensive hyaloclastite deposits running 200 km north from Vatnajokull, whereas Carswell (1983) mapped six sheet-like hyaloclastite units in the Plio-Pleistocene south of Myrdalsjokull.

This study describes and interpretes some very voluminous, extensive and stratigraphically conformable hyaloclastite sheets. They have probably been transported more than 50 km. The emplacement mechanism of these sheets adheres to a modified version of the flow hypothesis suggested by Walker and Blake (1966). The eruption model established is based on detailed mapping of internal lithofacies assemblages, vertical lithofacies sequences and lateral lithofacies relationships of the hyaloclastite units. The environmental reconstruction is based on comparisions made with the intercalated sedimentary rock horizons.

3. GEOLOGICAL SETTING AND REGIONAL TECTONICS

The Plio-Pleistocene sequence of the present study is situated in the Sida and Fljotshverfi districts just south of the eastern volcanic rift zone in southern Iceland (Fig. 1). This 500 square km area is limited to the northwest by the Postglacial volcanic fissure system of the Lakagigar

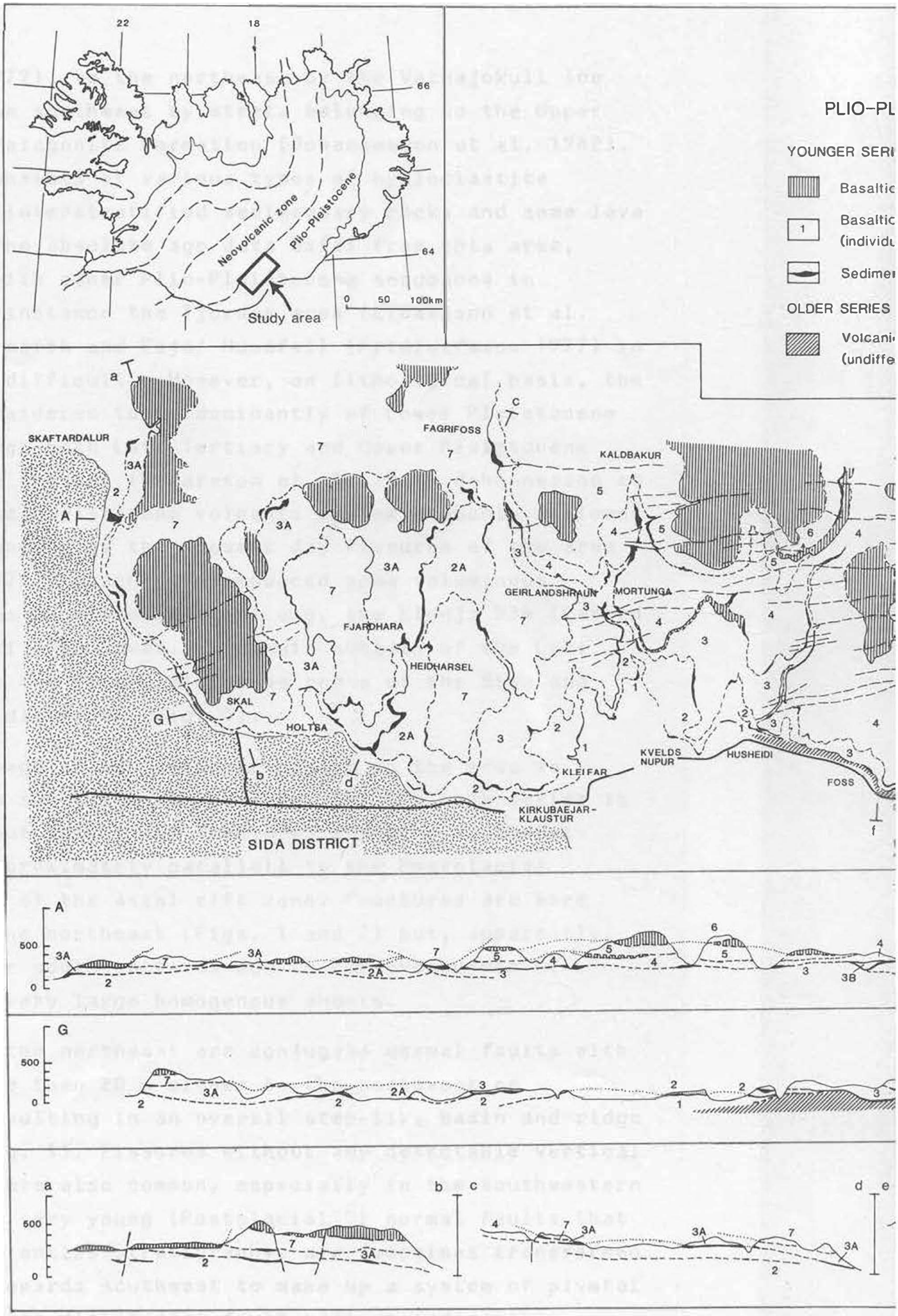
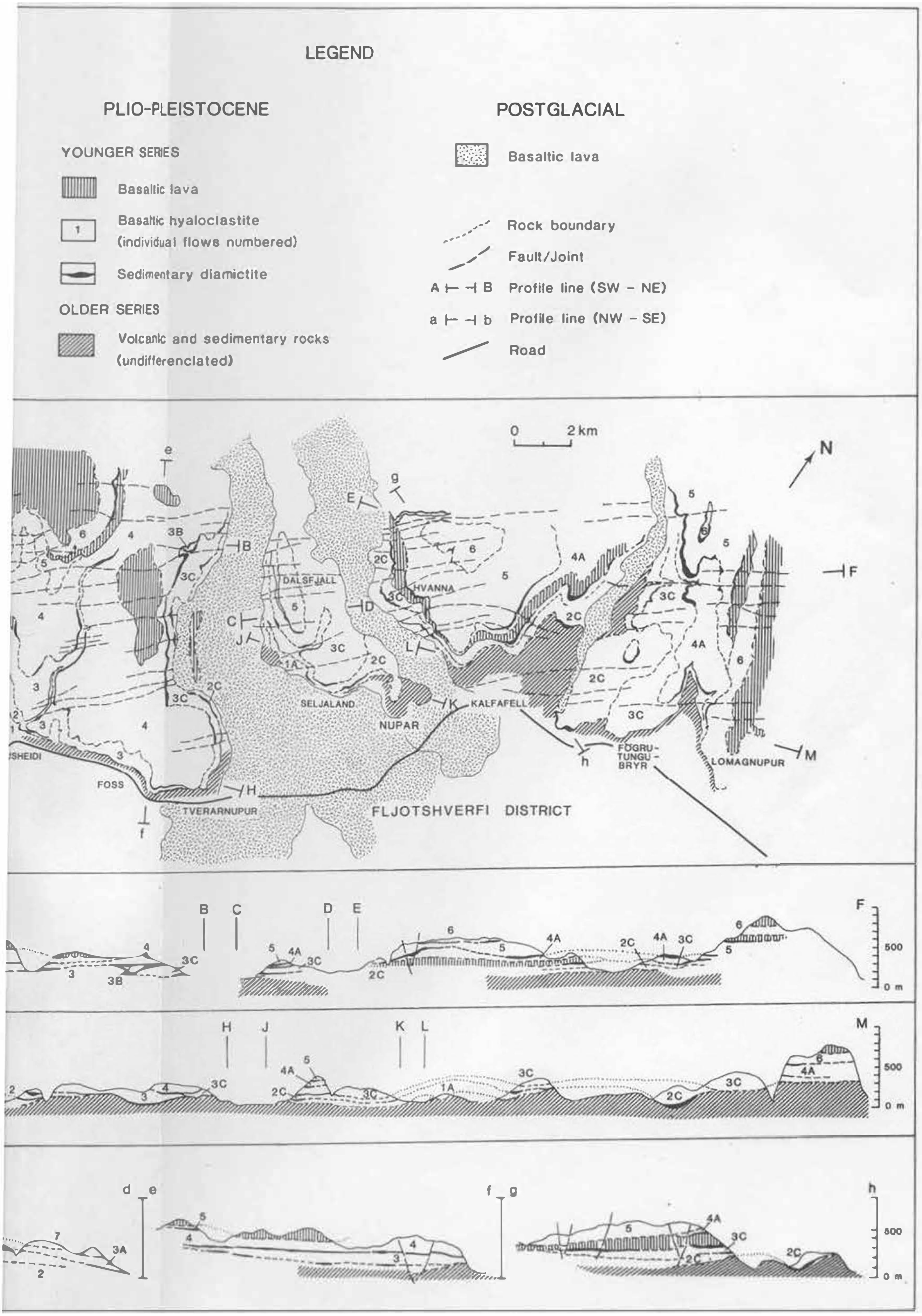


Fig. 1. Geological map and cross sections of the Sida and Fljotshverfi districts, southern Iceland.


LEGEND
PLIO-PLEISTOCENE
YOUNGER SERIES
Basaltic lava
Basaltic hyaloclastite
(individual flows numbered)
Sedimentary diamictite
OLDER SERIES
Volcanic and sedimentary rocks
(undifferenciated)
POSTGLACIAL
Basaltic lava
Rock boundary
Fault/Joint
A ⊢ ⊣ B Profile line (SW - NE)
a ⊢ ⊣ b Profile line (NW - SE)
Road
0 2 km
N
DALSFJALL
HVANNA
SELJALAND
NUPAR
KALFAFELL
FOGRU-
TUNGU-
BRYR
LOMAGNUPUR
FOSS
TVERARNUPUR
FLJOTSHVERFI DISTRICT
500
0 m

tshverfi

(Jakobsson 1979), to the northeast by the Vatnajokull ice cap and to the southwest by strata belonging to the Upper Pleistocene Palagonite Formation (Johannesson et al. 1982). The series consists of various types of hyaloclastite breccias and interstratified sedimentary rocks and some lava horizons. As no absolute age data exist from this area, correlation with other Plio-Pleistocene sequences in Iceland, for instance the Tjornes area (Einarsson et al. 1967) in the north and Esja/ Husafell (Fridleifsson 1977) in the west, is difficult. However, on lithological basis, the rocks are considered to be dominantly of Lower Pleistocene age even though both Late Tertiary and Upper Pleistocene strata may be present (Einarsson et al. 1967, Johannesson et al. 1982). The Pleistocene volcanic system probably followed the same alignment as the present day fissures of the area (Jakobsson 1979), which have produced some voluminous basaltic lavas in historic time, e.g. the Eldgja 934 (Larsen 1984) and Laki 1783 lavas. Two main outlets of the Laki lava transect the Plio-Pleistocene rocks of the Sida and Fljotshverfi districts (Fig. 1).

A striking image of the tectonic system in the area is obtained from satellite photos (Fig. 2). The rock series is cut by numerous joints and fractures trending southwest-northeast, approximately parallell to the Postglacial fissure swarm of the axial rift zone. Fractures are more abundant in the northeast (Figs. 1 and 2) but, apparently, they disappear southwestwards due to the overburden of younger (?), very large homogenous sheets.

Fractures in the northeast are conjugate normal faults with downthrow less than 20 m either to the northwest or southeast, resulting in an overall step-like basin and ridge structure (Fig. 1). Fissures without any detectable vertical displacement are also common, especially in the southwestern domains. Some very young (Postglacial ?) normal faults that cross-cut the entire stratigraphy, are sometimes transformed into joints towards southwest to make up a system of pivotal faults. The Plio-Pleistocene fault pattern indicates a

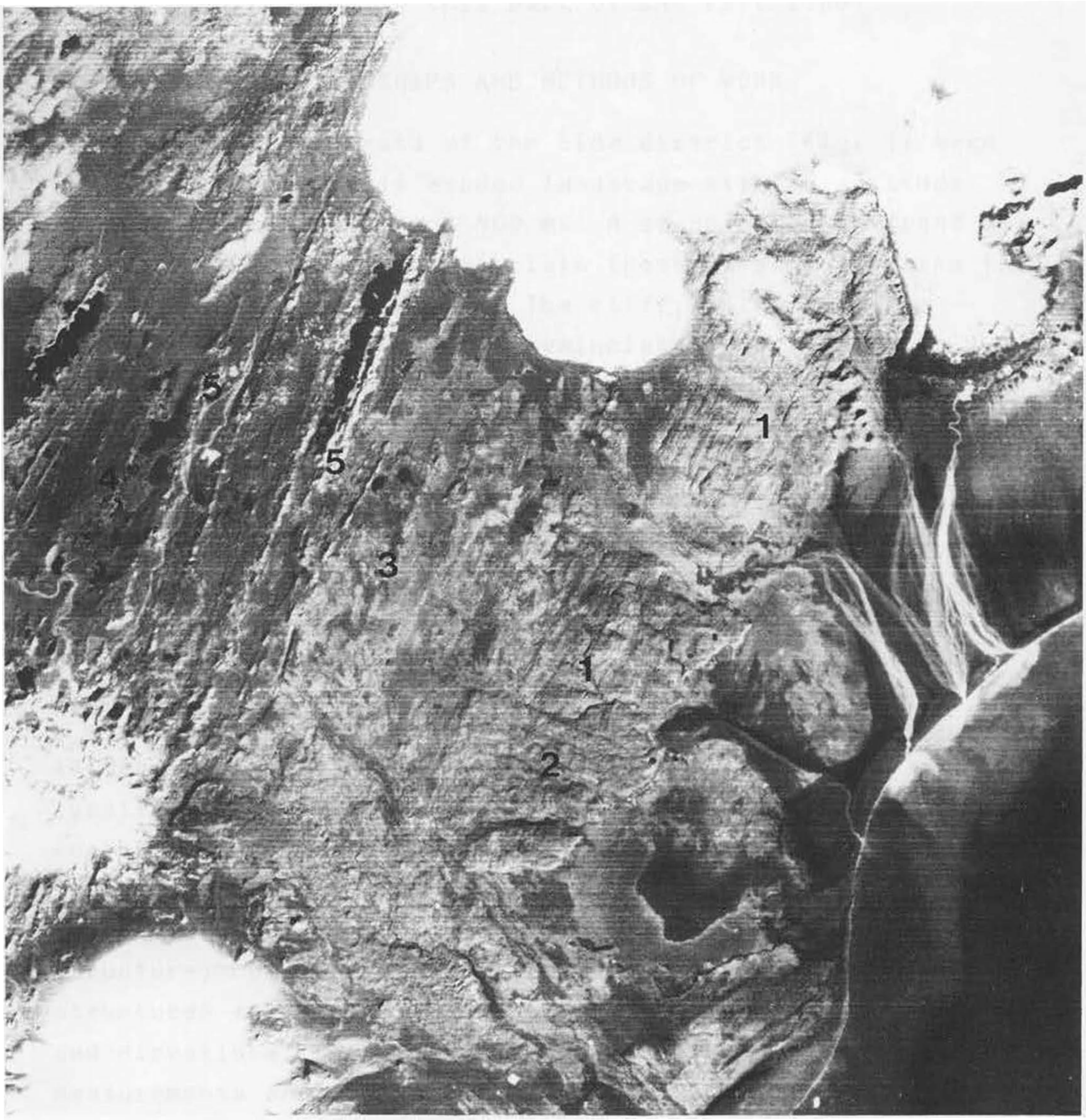


Fig. 2. NASA Earth Resources Technology Satellite image of the volcanic ridge-rift belt southwest of Vatnajokull. Plio-Pleistocene bed rock strata (1) with extensive NE-SW trending fractures are overlain by homogenous flow lobes in the SW (2). Crater rows of the present day Laki (3) and Veidivotn (4) fissure systems are clearly defined. Prominent Upper Pleistocene hyaloclastite ridges (5) are also seen.

component of lateral extensional movement in a direction northwest-southeast, corresponding well with the present spreading direction in this part of the rift zone.

4. FIELD RELATIONSHIPS AND METHODS OF WORK

Plio-Pleistocene strata of the Sida district (Fig. 1) make up a smooth, glacially eroded landscape with an altitude averaging between 100 and 400 m. A steep cliff in front of the present glacial outwash plain (coastal sandur) marks the former Postglacial sea-line. The cliff, which contains well-exposed outcrops of the hyaloclastite formation, extends southwest-northeastwards across the entire area. Further inland (northwestwards) the plateau is in general covered by thick soil/tephra layers, formed by Postglacial eruptions. It is cut by numerous deep (100 to 200 m) canyons that allow easy access to older rock strata. The Fljotshverfi district (Fig. 1) is characterized by steep mountains with relatively flat tops, up to 800 m, deeply cut by broad, U-shaped valleys during the Upper Pleistocene to Recent glacial period.

44 detailed sections have been measured from various localities in traverses both southwest-northeast and northwest-southeast (Fig. 3), in order to establish a stratigraphic framework and to evaluate vertical and lateral facies changes (Fig. 4). Bed thickness, lithology, internal structure, composition, grain size and volcano-sedimentary structures were noted, and whenever possible, bed contacts and directional features were outlined. Detailed grain size measurements and textural analyses were made at a few well-exposed outcrops.

5. GENERAL STRATIGRAPHY

Four generalized type stratigraphic columns are established in the mapped area (Fig. 5), on basis of the detailed sections outlined in Fig. 4. The Lomagnupur Section to the northeast, approximately 700 m thick, represents the most

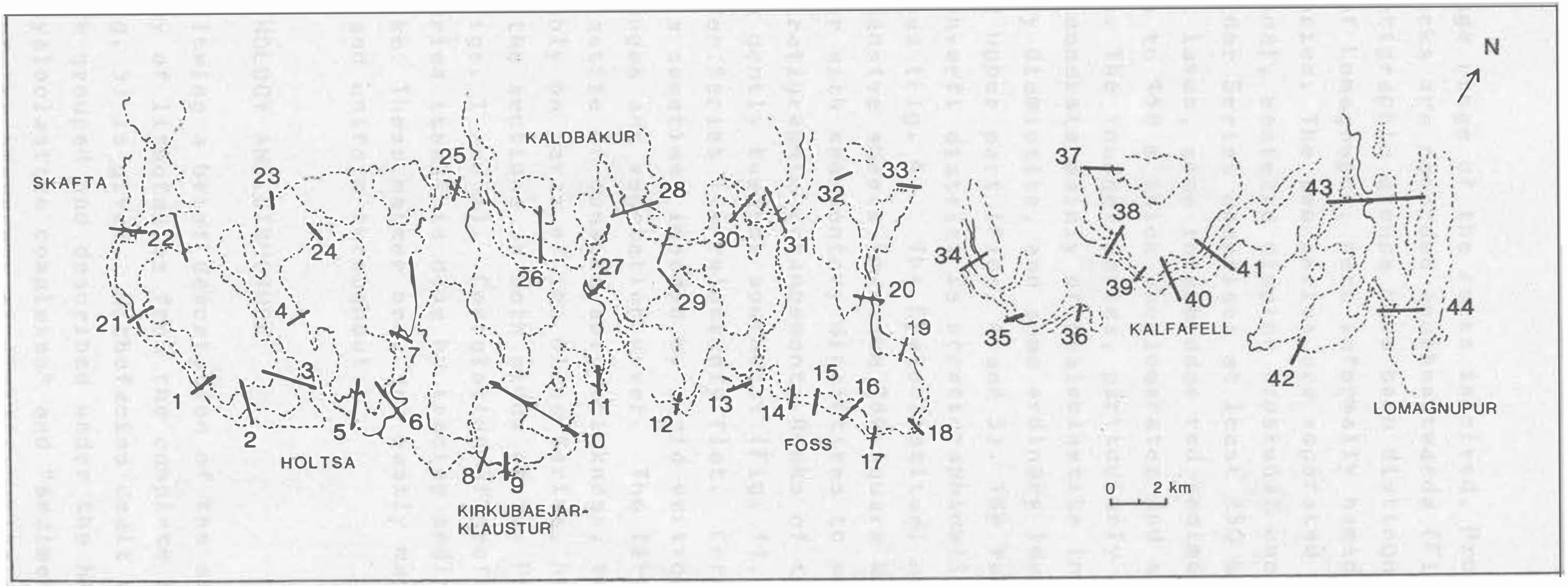


Fig. 3. Locality map of measered vertical sections.

complete age range of the rocks involved. Progressivly younger rocks are recorded southwestwards (Fig. 5). Two major stratigraphic groups have been distinguished from the sequence of Lomagnupur, here informally named the Older and Younger Series. The two series are separated from each other by a regional, westward dipping erosional unconformity (Fig. 5). The Older Series comprises at least 250 m of exposed horizontal lavas, some interbedded red sedimentary rocks and locally up to 150 m thick conglomerates and stratified sandstones. The Younger Series, particularly of the Sida district, consists mainly of hyaloclastite interbedded with sedimentary diamictite, and some ordinary lava flows in the middle and upper part (Figs. 4 and 5). The Younger Series of the Fljotshverfi district is stratigraphically more heterogenous (Fig. 4). The hyaloclastites, of both areas, define extensive sheets (maximum 280 square kilometres) that interfinger with sedimentary diamictites to make up a very complex stratigraphic arrangement. Rocks of the Younger Series dip gently towards southeast (Fig. 1), whereas lavas of the Older Series lie relatively flat. Correlation of sections is sometimes impeded by rapid vertical and lateral facies changes and vegetation cover. The first occurrence of hyaloclastite of considerably thickness, resting unconformably on lavas of the Older Series, has been used to correlate the sections on both sides of the Postglacial valleys (Figs. 1 and 4). Correlation of profiles within the Younger Series itself is done by tracing sedimentary layers along strike. These marker beds are easily mappable, very extensive and uniform throughout.

6. LITHOLOGY AND STRUCTURE

In the following a brief description of the structure and petrography of lithofacies from the complete stratigraphic scheme (Fig. 5) is given. Lithofacies dealt with in this account are grouped and described under the headings; "lava flows", "hyaloclastite complexes" and "sedimentary rocks". A summary of the hierarchy of classification schemes in use

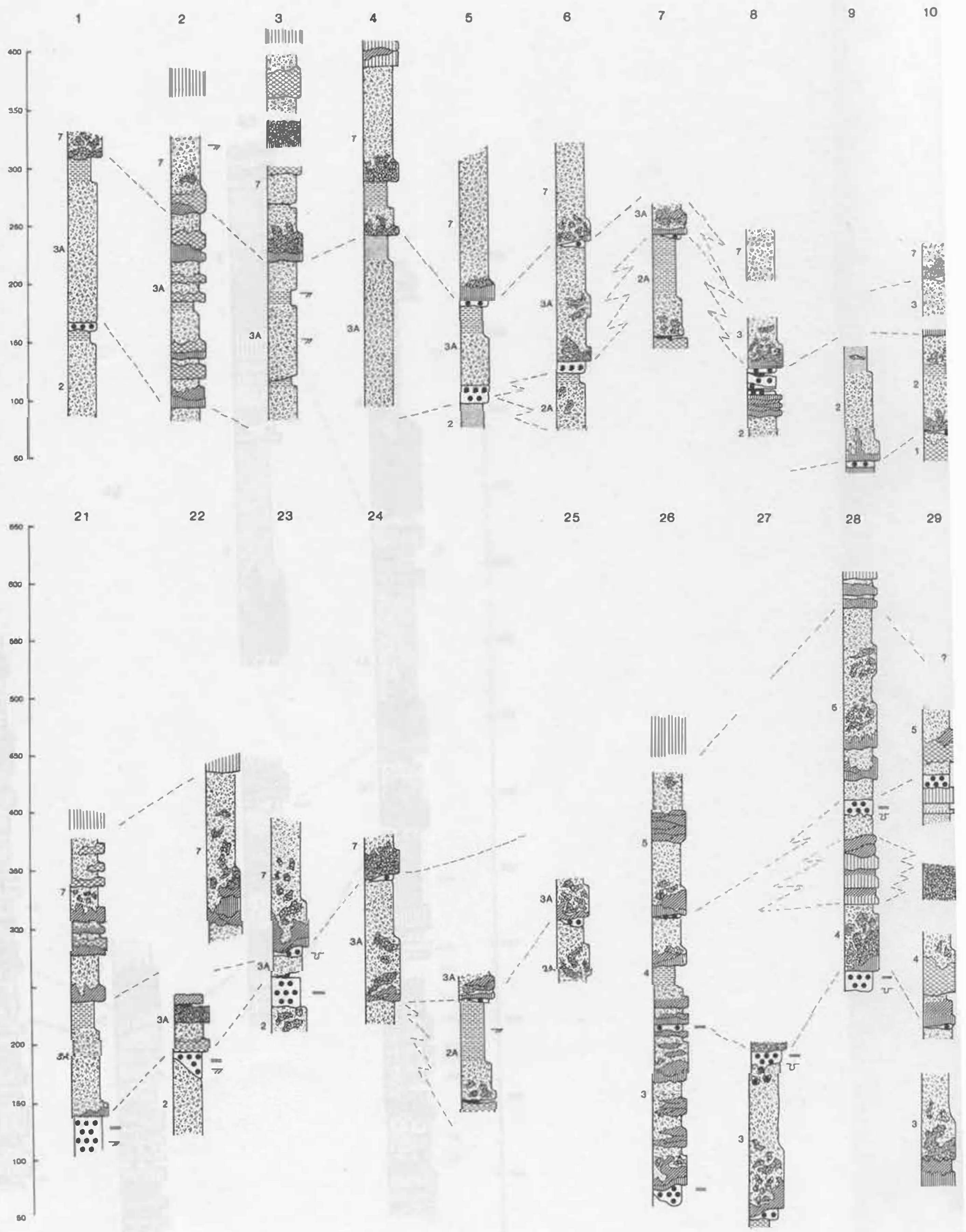
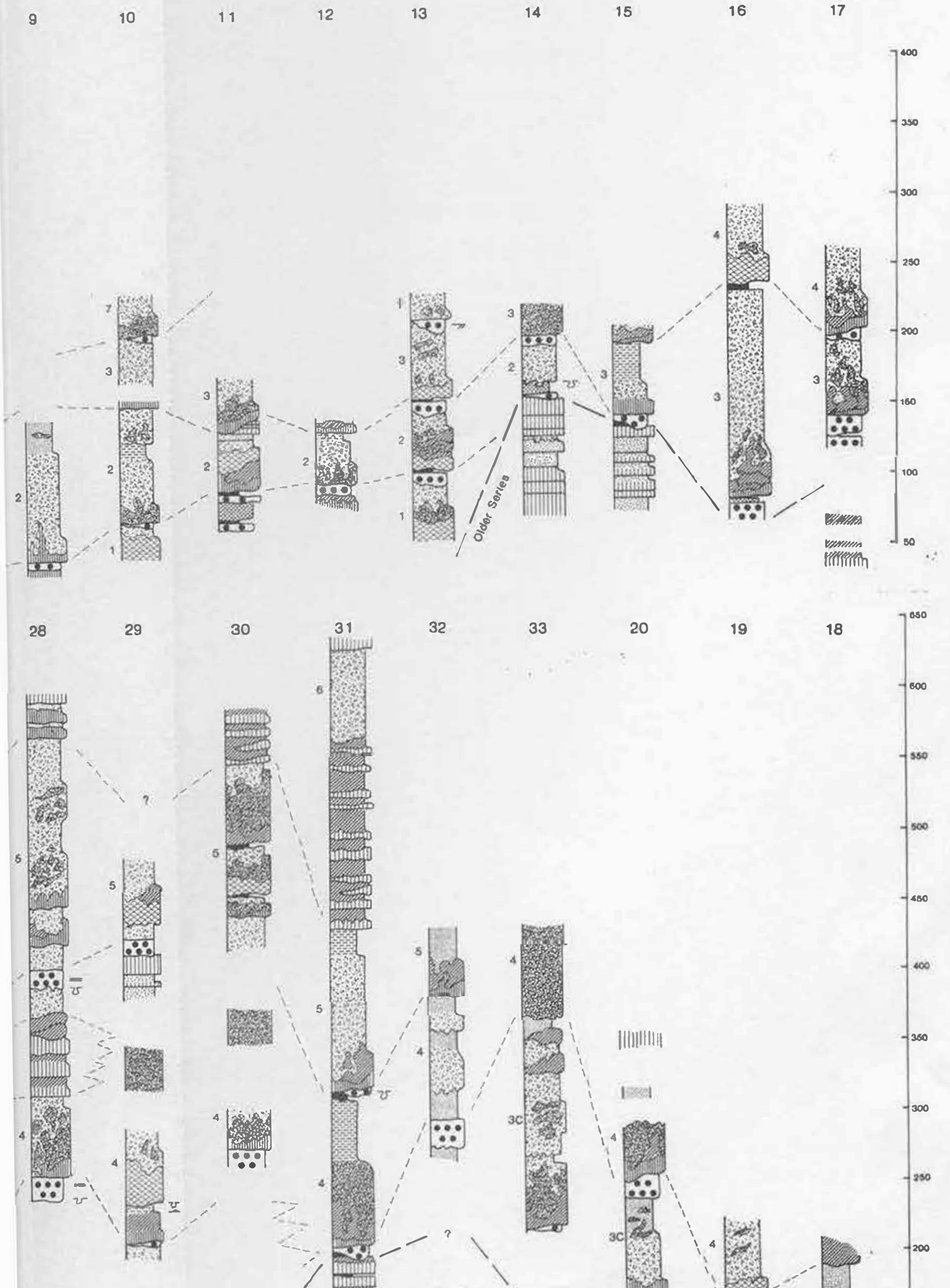


Fig. 4. Detailed measured sections through Plio-pleistocene strata. Legend symbols are shown in Fig. 5. Correlation of rock units is made by stippled lines. Number of individual flows is shown to the left of the lithology column, primary structures to the right. Decreasing width of the lithology column signifies increased decomposition of the basalt, and reduced grain size. Lateral position not to any scale. Height scale in metres above sea level.



nd

lumn,
signifies
ition

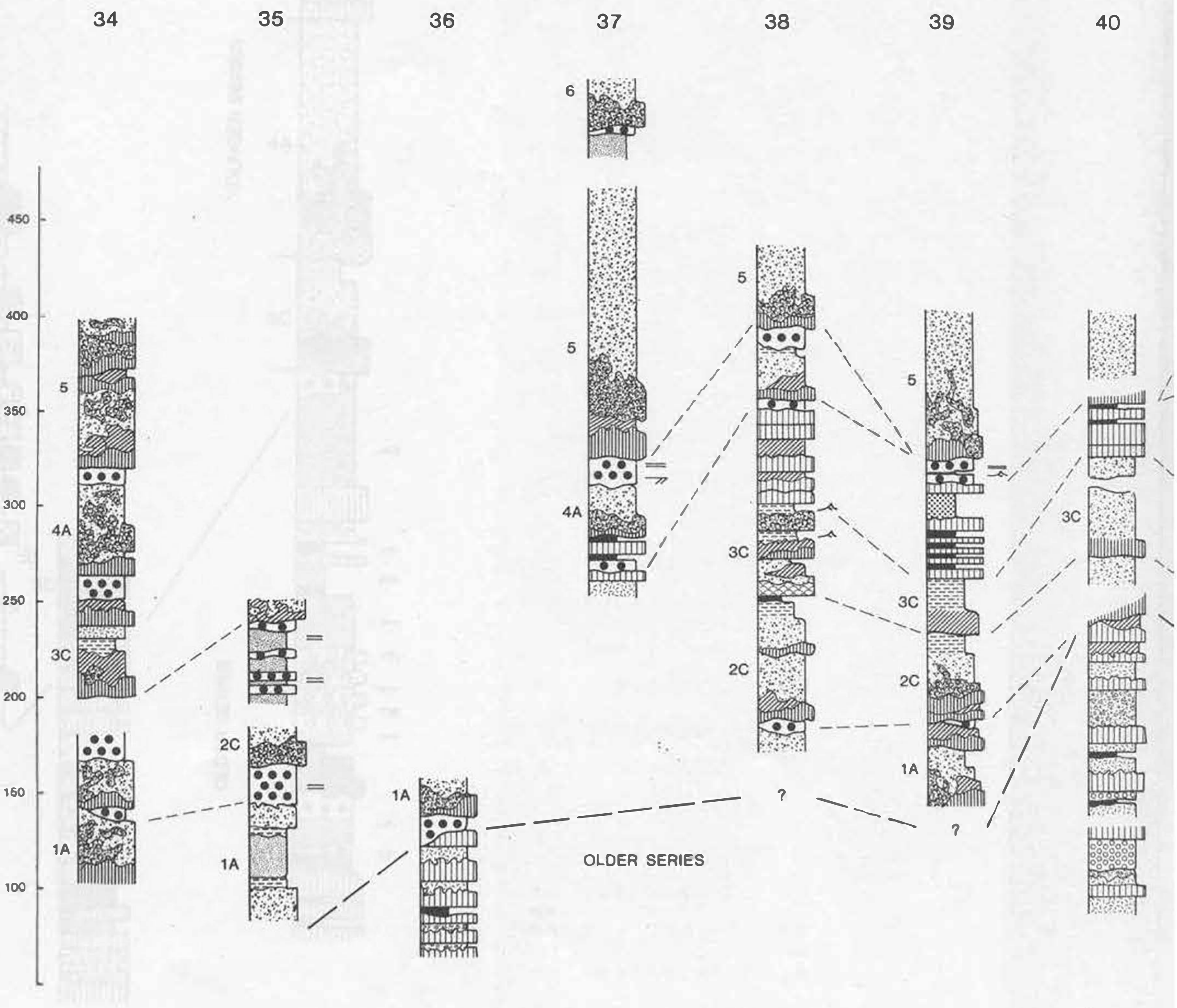
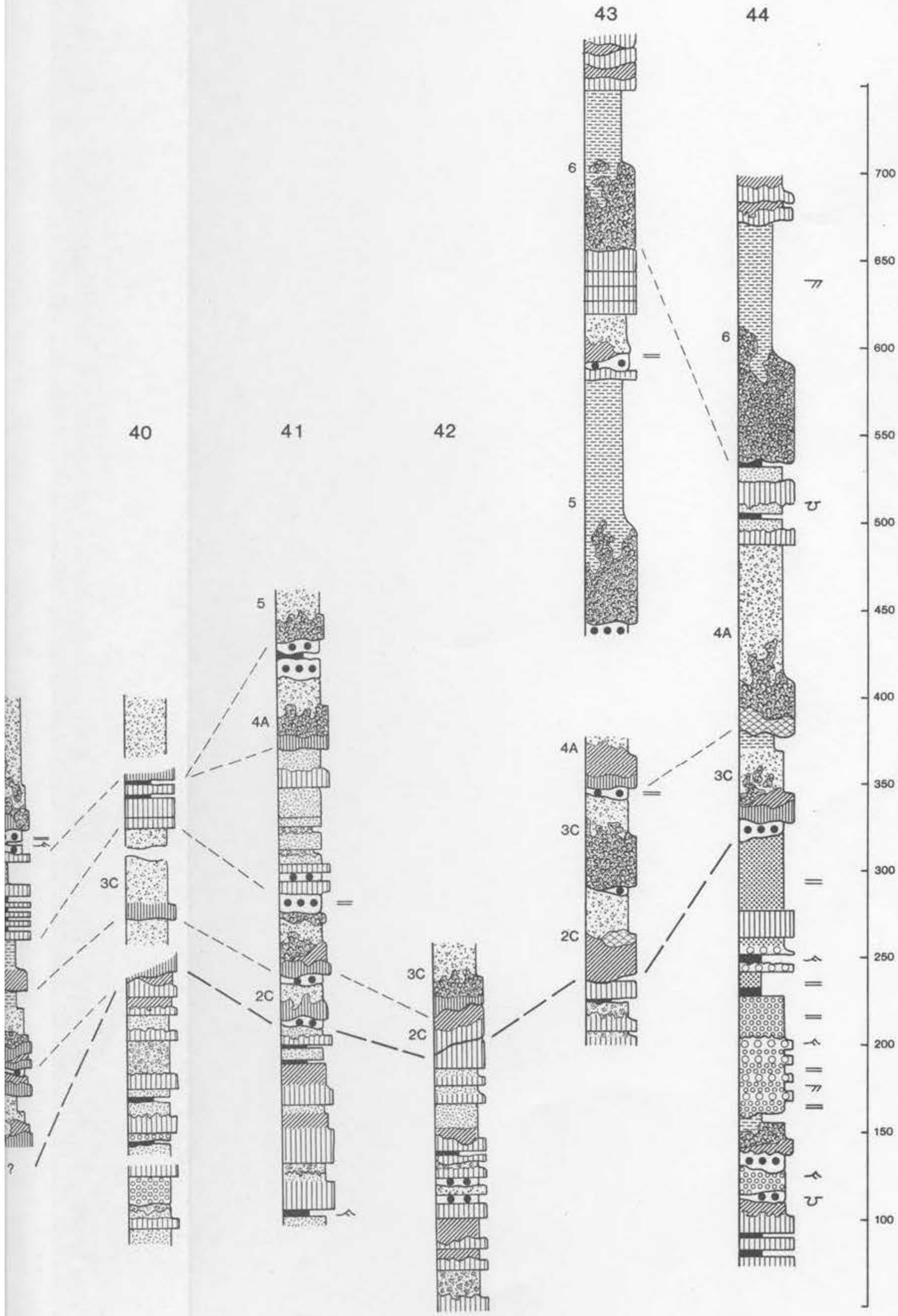


Fig. 4. (continued)


40
41
42
43
44
700
650
600
550
500
450
400
350
300
250
200
150
100
6
5
4A
3C
2C

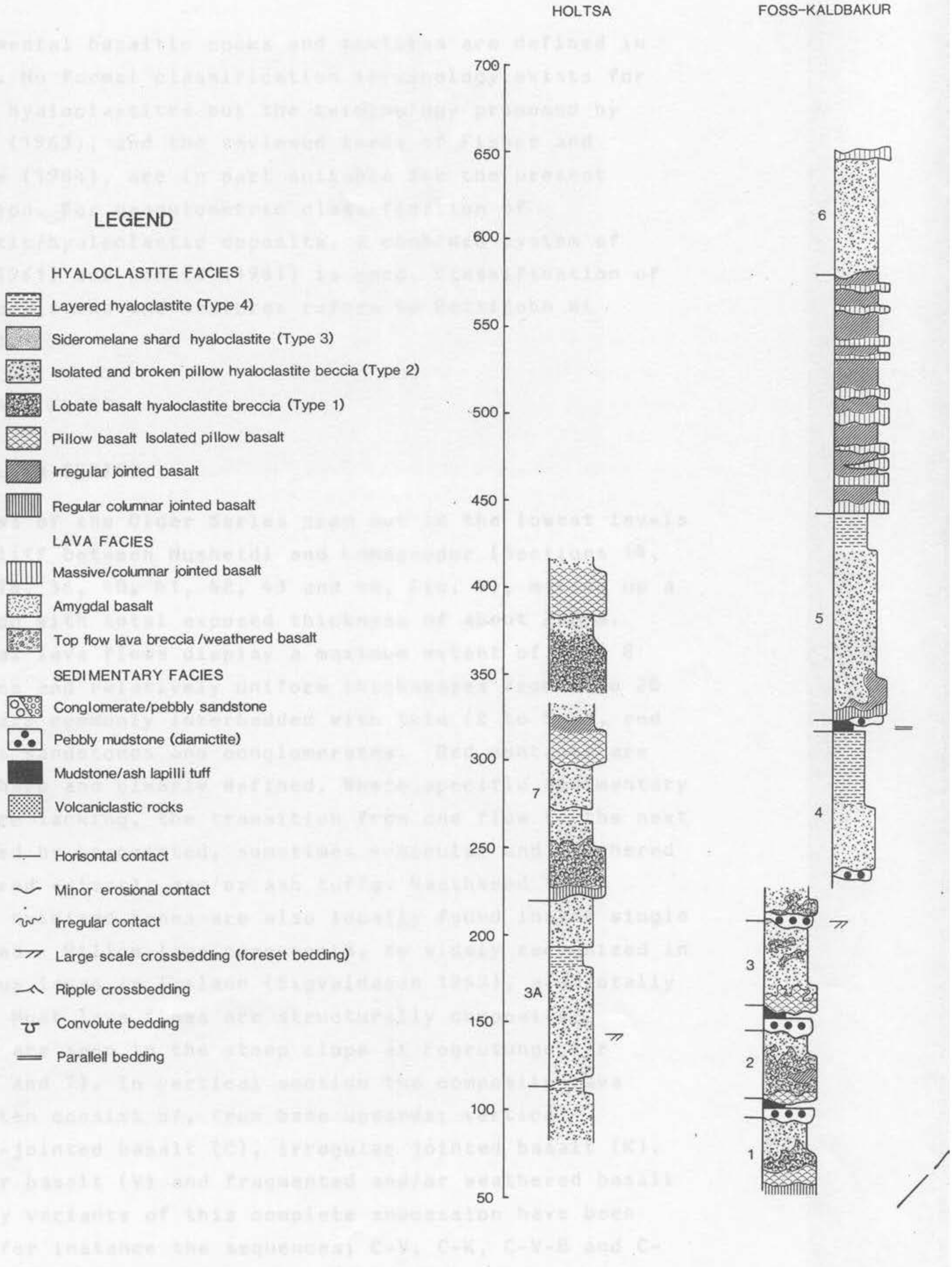
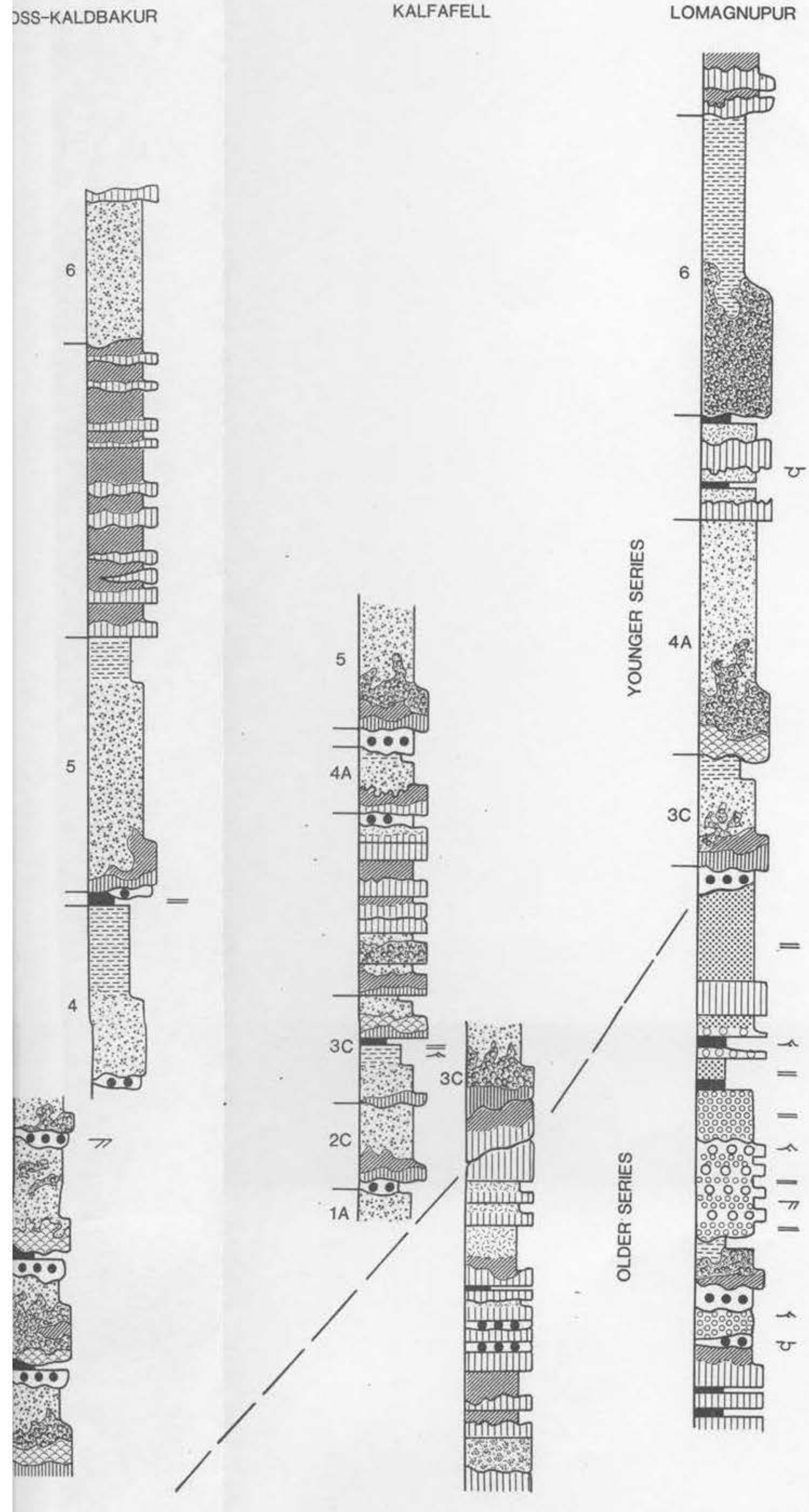


Fig. 5. Generalized stratigraphy scheme of the Sida and Fljotshverfi districts. For location of the sections, see Figs. 1 and 3.



fi districts.

for fragmental basaltic rocks and textures are defined in the text. No formal classification terminology exists for basaltic hyaloclastites but the terminology proposed by Carlisle (1963), and the reviewed terms of Fisher and Schmincke (1984), are in part suitable for the present description. For granulometric classification of pyroclastic/hyaloclastic deposits, a combined system of Fisher (1961) and Schmid (1981) is used. Classification of sedimentary rocks and textures refers to Pettijohn et al.(1972).

6.1. LAVA FLOWS

6.1-1. OLDER SERIES:

Lava flows of the Older Series crop out in the lowest levels of the cliff between Husheidi and Lomagnupur (Sections 14, 15, 17, 18, 36, 40, 41, 42, 43 and 44, Fig. 4), making up a succession with total exposed thickness of about 250 m. Individual lava flows display a maximum extent of 15 X 8 kilometres and relatively uniform thicknesses from 2 to 20 m. They are commonly interbedded with thin (2 to 5 m), red mudstones/sandstones and conglomerates. Bed contacts are always sharp and clearly defined. Where specific sedimentary layers are lacking, the transition from one flow to the next is defined by brecciated, sometimes vesicular and weathered basalt (red colors), and/or ash tuffs. Weathered and strongly oxidized zones are also locally found inside single lava flows. Pillow lava components, so widely recognized in subaqueous lavas in Iceland (Sigvaldason 1968), are totally absent. Most lava flows are structurally composite. Examples are seen in the steep slope at Fogrutungubryr (Figs. 6 and 7). In vertical section the composite lava flows often consist of, from base upwards; vertical columnar-jointed basalt (C), irregular jointed basalt (K), vesicular basalt (V) and fragmented and/or weathered basalt (B). Many variants of this complete succession have been mapped, for instance the sequences; C-V, C-K, C-V-B and C-

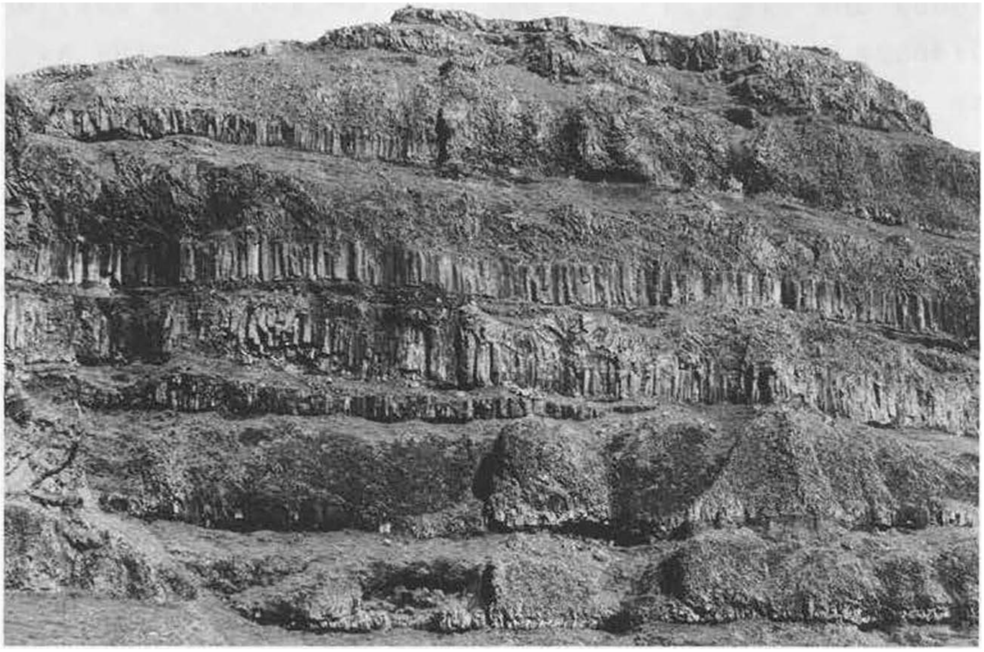


Fig. 6. Basalt lavas of the Older Series exposed in the cliff at Fogrutungubryr. Note regular vertical columnar jointing at the bases and irregular and rosett-shaped kubbaberg columns in the upper half of the flows. Height of cliff about 150 m.

Fig. 7. Flat-lying kubbaberg lava flows of the Older Series overlain by massive hyaloclastite units of the Younger Series. From a locality 500 m NE of Fogrutungubryr.

K-V-B (see Sections 40, 41, and 42, Fig. 4). The succession C-K, in which a lower zone of regular columnar structures is sharply overlain by hackly jointed columnar basalt, is a very abundant structure association of lavas of the Palagonite Formation in Iceland, known as "kubbaberg" (Sigvaldason 1968, Saemundsson 1970). The type locality for study of such two-tired columnar basalt lavas is Dverghamrar, just south of Foss (Fig. 1). A common feature of the kubbaberg lava associations is an upward fanning and curvature of the columns to produce rosett-shaped arrangements (Fig. 6). Locally small pockets of hyaloclastite and basalt breccia may be present at the very top of such irregularly jointed flows, assuming that the lava was partly fragmented during emplacement and/or the initial cooling phase. More extreme effects of fragmentation are documented from the Younger Series (see below).

Another widespread vertical succession of the composite lava flows is that in which vesicular basalt (V), rests either on columnar (C) or kubbaberg (K) facies and is gradually overlain by very thin (less than 50 cm), porous basalt breccia (B) (see Sections 14, 40, 41,42 and 43, Fig. 4). Aligned vesicles, up to 5 cm in size, occur at the flow tops, and usually they are filled with zeolites. Vesicular flow layers are sometimes present in the centres of some flows. The frequency and size of vesicles then typically increase upwards, particularly where basalt breccias are present at the top. Very thick (almost 15 m) breccias are observed in amygdal flows in the lower levels at Kalfafell (Section 40) and Raudaberg (Section 42). The breccia consists of multiple layers of alternating blocky (less than 50 cm)-and lapilli-sized basalt fragments, with angular shapes and a highly porous composition. The surrounding matrix (less than 10 volume percent) consists essentially of finely crushed vitric particles and secondary zeolite. The gradual upward transition from homogenous basalt into amygdal basalt and basalt breccias at the top, proves that the breccias are formed by disintegration of basalt in the same flow.

A change from grey aphyric basalt in the lower parts of some flows, to extremely porous and brecciated red basalt at the top (see Sections 40, 41 and 42, Fig. 4) is considered to be the result of subaerial weathering. These red lavas are nearly always interbedded with red sedimentary mudstones. The red coloration is due to hematite coatings on mineral grains.

Simple lava flows of the Older Series are either very coarse columnar jointed (up to 2 m in diameter), massive and/or vesicular throughout. The thickness of the flows are generally less than 20 m and corresponds well with the thickness of the lower massive part of kubbaberg flows, suggesting that they may be eroded remnants of the composite flows.

Petrographically, lavas of the Older Series are dark, bluish grey in color and vary from weakly porphyric to almost aphyric. Thin-section studies reveal fresh euhedral plagioclase and olivine phenocrysts (5 mm) and locally clinopyroxene in a medium/fine-grained intersertal texture. The modal content of total phenocrysts may be up to 10 %. Lath-shaped plagioclase (15-30%) and crushed and corroded olivine(5-20%) and pyroxene (5-15%) are the main components of the matrix. An opaque mass (10-30%), locally stained with brown-colored iron oxide, infills the remaining space. The average grain size normally increases towards the centre of individual flows. The vesicular flows contain zeolites, of which chabazite is by far the most common species together with a green opal. The frequency of vesicles and most intensive zeolitization appear at the surfaces and particularly within the top flow breccias, due to higher permeability for circulating fluids.

6.1-2. YOUNGER SERIES:

Lava flows are less frequent in the Younger Series, but occur at some stratigraphic intervals. In the Fljotshverfi district lavas make up complex interbeds between the hyaloclastites, and displaying internal structure composition partly similar to those found in the Older Series (see Sections 38, 39 and 41, Fig. 4). In contrast, lava flows of the Sida district are relatively homogenous, somewhat thicker and laterally less extensive. Mappable flows here occupy three distinct levels (Figs. 1 and 4). The lowermost is exposed between 170 and 200 m.a.s., south of Mortunga (Fig. 1). The next occurs between 440 and 570 m.a.s. at Kaldbakur (Figs. 4 and 5, Section 3), comprising about ten lens-shaped kubbaberg flow units, approximately 12 m in thickness, and grading laterally into hyaloclastite. Finally, a very extensive flow, at least 40 m thick, covers the highest exposed levels of the stratigraphic division (Fig. 5), and is represented by several isolated outcrops (Fig. 1). It probably constitutes no more than one single flow unit, as seen from the uniform composition and absence of internal surfaces and structure variations. This basalt is olivin-plagioclase-phyric and relatively coarse-grained, and typically contains large empty vesicles up to 3 cm in size.

6.1-3. INTERPRETATION:

The lateral persistence, uniform thickness and structural continuity of lavas in the Older Series indicate deposition as broad lava plains extruded from fissures and spread across a flat plateau landscape more or less similar to that of the Tertiary series in Iceland. A direct comparision of flow mechanism is, however, unlikely since kubbaberg basalts are generally absent in the Tertiary. The origin of two-tired columnar jointing (kubbaberg) zones in basaltic lavas is debated (Waters 1960, Saemundsson 1967). Three models have been proposed; one is by flow in a subaqueous environment (Sigvaldason 1968), another is by temporary overfloodings of water upon a still fluid subaerial lava, resulting in sudden thermal cooling and incipient lava fragmentation (Waters 1960, Saemundsson 1967). The third explanation is when a thick fluid lava, unrelated to environment, suffers irregular stress distribution during emplacement, for instance faster movement in the upper part and slower movement at the base (Waters 1960, Spry 1962, Williams and Mc Birney 1979).

Lava flows of the Older Series show some evidence of differential movements, as documented by the fanning columns and frequent alignment of vesicles in various parts of the flows. The commonly observed flow breccias and local hyaloclastites on top of the flows, suggest incipient surface brecciation due to thermal cooling and quenching with water. Similar top flow breccias, termed hyalotuff-breccias by Dimroth et al. (1978), have been described from very gas-rich lavas running into water (Jones 1969). A complete subaqueous environment of deposition of lavas from the Older Series is, however, disregarded from the presence of subaerial weathering and oxidation effects, general lack of pillow lava, hyaloclastite and other water-soaked immersed structures, and by the common interbedding with terrestrial sedimentary rocks (see below). Similarly, there is no evidence that explosive volcanic activity played any important role during emplacement of these flows, as would probably have been expected in a shallow subaqueous environment. The high proportion of amygdules, however, suggests that the lavas were originally quite gas rich. This, combined with the fact that kubbaberg structures and brecciated flow tops are relatively restricted along strike, favors the model of formation of these facies by temporary overfloodings of water on the still hot lava surface, for instance in fluvial channels and gullies (Saemundsson 1967).

The kubbaberg lava facies found in the Younger Series were probably formed in a subaqueous environment, as indicated from the common lateral grading into hyaloclastite facies and general absence of interbedded subaerial sedimentary rocks (see below). On the other hand, ice-free or subaerial regions could have locally existed, in the Fljotshverfi district, where some red oxidized lava breccias have been observed (Sections 39, 40 and 41, Fig. 4).

6.2. HYALOCLASTITE COMPLEXES

6.2-2. NOMENCLATURE:

The term hyaloclastite was introduced by Rittmann (1960) for rocks composed of glassy particles produced by non-explosive disintegration of pillow lava. In Iceland this term applies to brecciated rocks of the subglacial Palagonite System formed not only by granulation of lava in contact with ice, but also by explosive phases of underwater eruptions. The definition we use follows Fisher and Schmincke (1984) to include all vitroclastic breccias produced by explosive and non-explosive interraction of water and hot magma. "Pillow lava" (P) of this study, defines close-packed pillow structured basalt with less than 10 % hyaloclastite matrix." Isolated pillow basalt" (IPB) designates a rock containing whole, isolated pillows with matrix up to 20 %. The general term "hyaloclastite breccia" (HB), with a matrix content exceeding 20 %, is used to cover all observed varieties of hyaloclastite. Noe-Nygaard (1940) employed the terms "globular basalt" and "basalt globe breccia" to designate pillow lava and hyaloclastite with globular basalt inclusions. The term globular basalt was originally used by Thoroddsen (1925) and Peacock (1926) to describe pillow-formed masses considered intrusive in origin. In recent litterature the name "lobate basalt" is used for subcircular basalt bodies enclosed in hyaloclastite (Gronvold 1972, Saemundson 1972). Furnes and Fridleifsson (1980) refered to "lobes" in silicic hyaloclastite formed by spall off from a magma that were formerly intruded into the hyaloclastite. Since no formal accepted nomenclature for lobe formation and occurrences in hyaloclastite has been published, we prefer the name "lobate basalt hyaloclastite breccia" (LO) for hyaloclastite containing lobate basalt inclusions. With increasing matrix proportion of the hyaloclastites, they are classified as "isolated and broken pillow hyaloclastite breccia" (IBHB) and "sideromelane shard hyaloclastite" (SSH), respectively. The descriptive term "hyaloclastite tuff" (LH) is applied to fine-grained layered hyaloclastite deposits resembling hyalotuffs, which has a strictly genetic significance (Honnorez and Kirst 1975, Dimroth 1977, Dimroth et al. 1978).

6.2-2. FIELD DESCRIPTION:

Fourteen separate hyaloclastite units, or complexes, have been identified and mapped in the Younger Series (Fig. 1 and Table 1). On the map (Fig. 1) and detailed logs (Fig. 4) these are numbered (Hy-1,2 etc.) regarding to age, and with units in the same lateral position, but different region, given the subscript A, B, C etc. The thickness of each complex varies between 20 and 250 m, averaging 70 m. Maximum areal extents of nearly 300 km2 have been mapped (Table 1). Volumes based on minimum, average and maximum exposed thicknesses are in the range of 0.1 to 31.4 cubic kilometres (Table 1). Cross-sections northwest-southeast reveal consistent dip of the units towards southeast, with an average dip angle approximating 10 degrees (Figs. 1 and 8). In cross section northeast-southwest (Figs. 1, 4 and 9) many complexes display a sheeted, planar layer-cake geometry, others appear rather lenticular and wedge-shaped (the complexes 2A and 5, Figs. 1 and 4). A notable feature observed in this direction is a thickness increase where the rocks infill gullies in the substratum, while their upper boundaries are quite flat (Figs. 4 and 9). Topographically controlled thicknesses are most common in the stratigraphic lower levels where the hyaloclastites rest on the highly eroded Older Series (Figs. 1 and 10). Impressive examples are seen at Husheidi, Foss and Raudaberg (Fig. 10). At Husheidi three individual sheets fill up the paleovalley at different positions and wedge out towards the valley side. At Raudaberg the depression is filled first by a 40 m thick pile of sediment and then overlain by hyaloclastite (Fig. 10). Erosion of the Older Series is observed to be some 200

Flow	Minimum thickness	Maximum thickness	Average thickness	Area (km2)	Volume (km3) Min.	Aver.	Max.
1			35	50		1.8	
1A	15	60	35	80	1.2	2.8	4.8
2	30	100	60	210	6.3	12.6	21.0
2A	10	100	50	40	0.4	2.0	4.0
2C	20	70	45	105	2.1	4.7	7.4
3	25	170	97	182	4.5	17.6	30.8
3A	70	220	145	125	9.0	18.3	27.7
3B		50	25	5		0.1	
3C	20	75	47	250	5.0	11.8	18.7
4	25	150	88	280	7.0	24.5	42.0
4A	20	110	65	160	3.2	10.4	17.6
5	90	150	120	240	21.6	28.8	36.0
6	45	140	92	108	4.9	9.9	15.0
7	80	200	140	224	18.0	31.4	44.8

Table 1. Volume estimates of individual hyaloclastite flow units based on mapped areal and observed maximum and minimum thicknesses.

Fig. 8. Southeasterly view of the gently dipping, bedded diamictite surface separating Hy-1 and 2 at Kleifar.

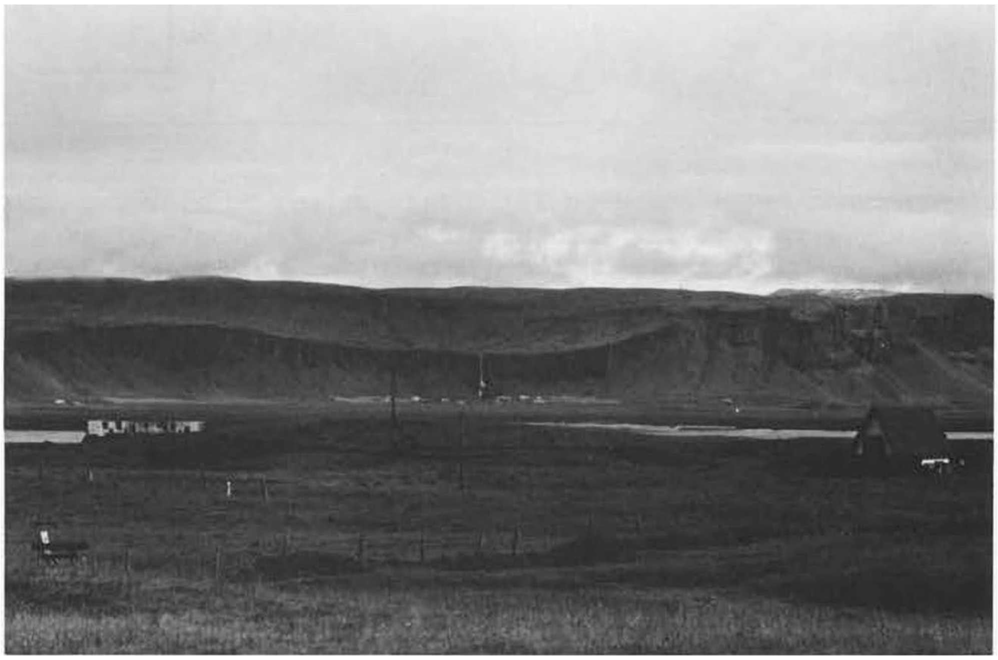


Fig. 9. Wedge-shaped hyaloclastite (Hy-3) at Foss, cut in direction SW-NE. Note thickness increase within the depression just above the farm, and the almost flat-lying bed overlying the hyaloclastite.

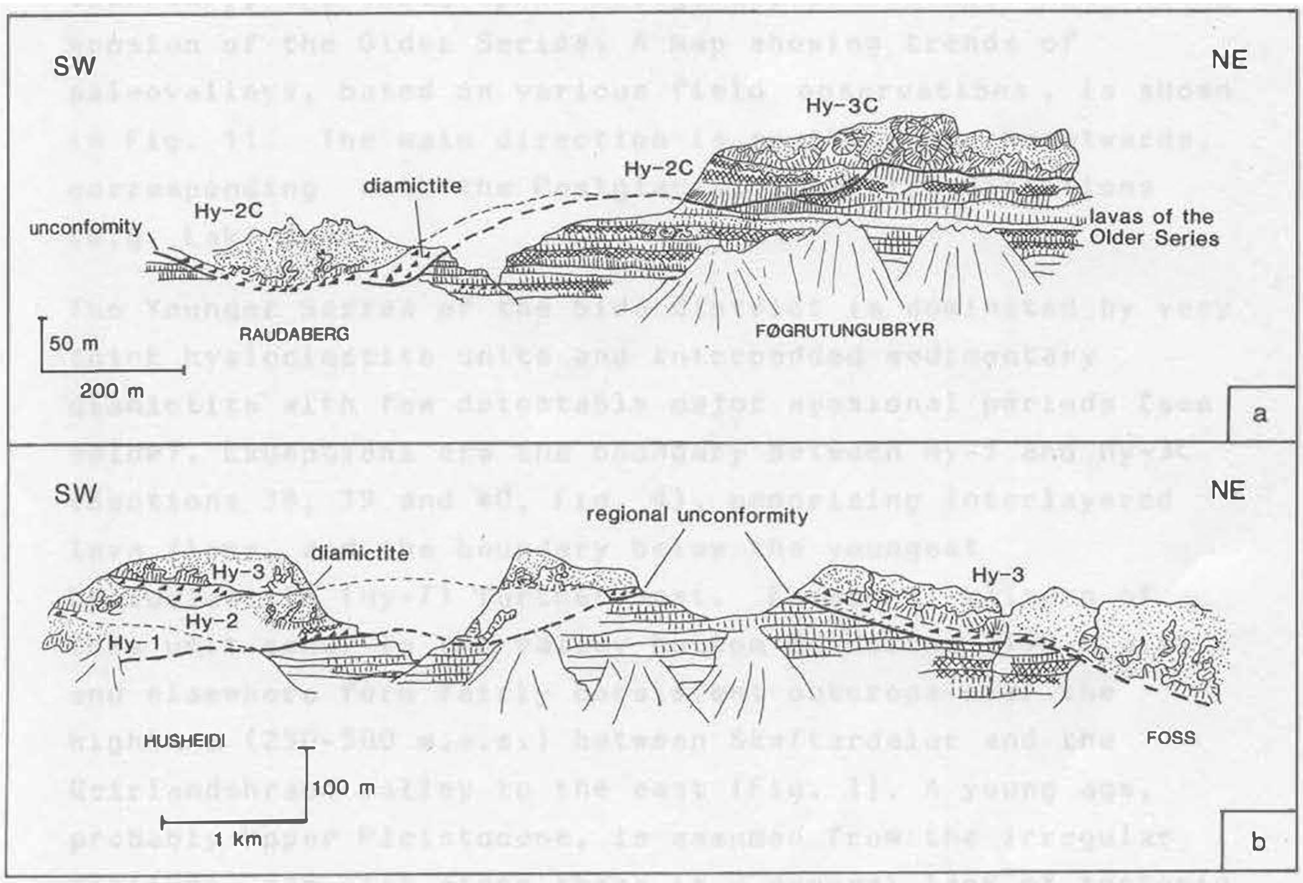


Fig. 10. Drawings of the coastal cliff between, a. Raudaberg-Nupstadir, and b. Husheidi-Foss, showing the irregular paleotopography of the major unconformity, separating the Older and Younger Series. Note infill of sedimentary diamictite in the depression at Raudaberg, and the wedging out of Hy-1 and 2 towards northeast, at Husheidi.

m at that locality. Thickness increases confined to paleovalleys are also noticed higher up in the Younger Series, but these are much smaller and probably reflect, secondary, the topographic irregularities established during erosion of the Older Series. A map showing trends of paleovalleys, based on various field observations , is shown in Fig. 11. The main direction is south to southeastwards, corresponding with the Postglacial lava flow directions (e.g. Laki lava).

The Younger Series of the Sida district is dominated by very thick hyaloclastite units and interbedded sedimentary diamictite with few detectable major erosional periods (see below). Exceptions are the boundary between Hy-5 and Hy-3C (Sections 38, 39 and 40, Fig. 4), comprising interlayered lava flows, and the boundary below the youngest hyaloclastite (Hy-7) further west. Erosional klippen of this unit occur in the valley bottom at Holtsa (150 m.a.s) and elsewhere form fairly consistent outcrops over the highland (250-500 m.a.s.) between Skaftardalur and the Geirlandshraun valley to the east (Fig. 1). A young age, probably Upper Pleistocene, is assumed from the irregular altitude, and also since there is a general lack of tectonic fractures so commonly found in strata further east (Fig. 2).

6.2-3. VERTICAL STRUCTURAL ORGANIZATION:

The internal structure and composition of the hyaloclastite complexes are, overall, amazingly regular, but variations do exist (Fig. 4) on smaller scale. The use of the term "complex" designated in this text, reflects this heterogenity, as well as the probability of finding more than one flow units within the same unit. A detailed vertical facies analysis (De Raaf et al. 1964) has been undertaken for the hyaloclastite complexes (Fig. 12) by compiling all measured profiles (Fig. 4). This reveals a generalized, vertical succession of individual hyaloclastite flow units, shown in Fig. 12. It starts with a massive, columnar jointed basalt (C), grading upwards into either;

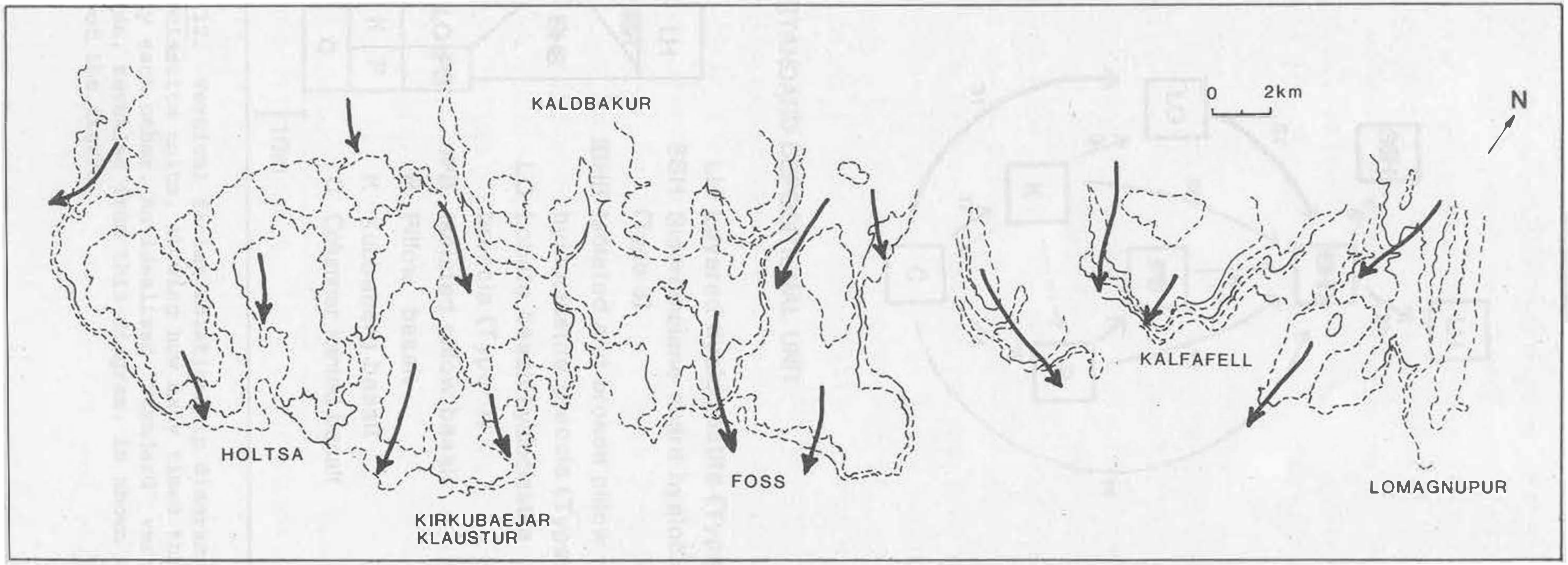


Fig. 11. Map showing axes of paleovalleys and main flow directions of the hyaloclastite units.

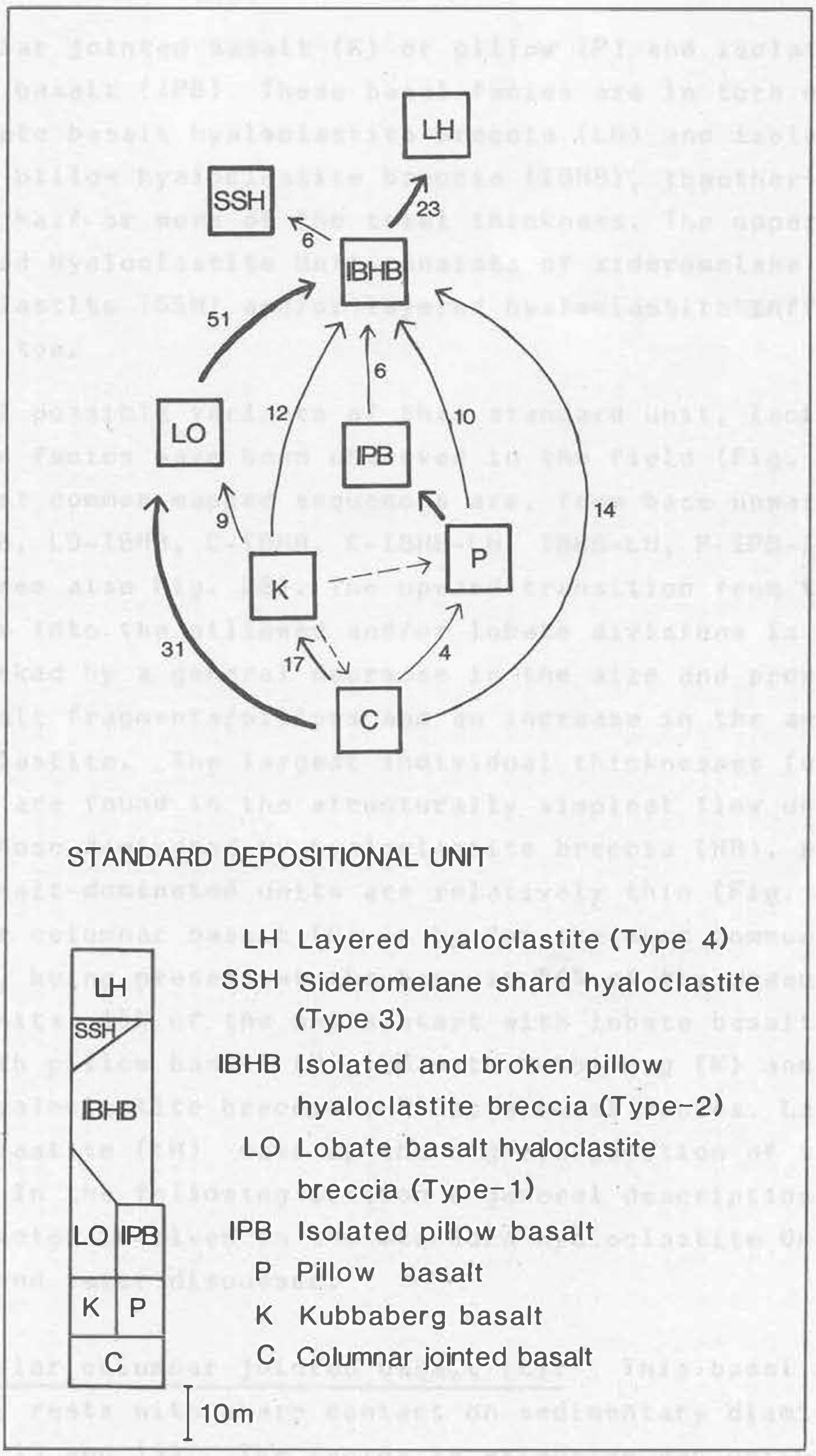


Fig. 12. Vertical facies relationship diagram for composite hyaloclastite units, showing how many times the different facies overly each other. An idealized "Standard" vertical cross section, revealed from this diagram, is shown in the lower half of the figure.

irregular jointed basalt (K) or pillow (P) and isolated pillow basalt (IPB). These basal facies are in turn overlain by lobate basalt hyaloclastite breccia (LO) and isolated and broken pillow hyaloclastite breccia (IBHB), together making up one half or more of the total thickness. The uppermost Standard Hyaloclastite Unit consists of sideromelane shard hyaloclastite (SSH) and/or layered hyaloclastite tuff (LH) at the top.

Several possible variants of this standard unit, lacking one or more facies have been observed in the field (Fig. 12). The most common mapped sequences are, from base upwards: C-LO-IBHB, LO-IBHB, C-IBHB, C-IBHB-LH, IBHB-LH, P-IPB-IBHB and IBHB (see also Fig. 28). The upward transition from the massive into the pillowed and/or lobate divisions is gradual and marked by a general decrease in the size and proportion of basalt fragments/pillows and an increase in the amount of hyaloclastite. The largest individual thicknesses (up to 120 m) are found in the structurally simplest flow units, e.g. those dominated by hyaloclastite breccia (HB), whereas the basalt-dominated units are relatively thin (Fig. 4). Regular columnar basalt (C) is by far the most common basal facies, being present at the base in 56% of the measured flow units. 15% of the units start with lobate basalt (LO), 12% with pillow basalt (P), 4% with kubbaberg (K) and 12% with hyaloclastite breccia (HB) as a basal facies. Layered hyaloclastite (LH) make up the highest position of the units. In the following section a general description of the lithofacies involved in the Standard Hyaloclastite Unit is given and later discussed.

a. Regular columnar jointed basalt (C): This basal member usually rests with sharp contact on sedimentary diamictites (Figs. 13 and 15). The facies is extensive and uniform, and can be traced several kilometres along strike. The thickness varies from 5 to 30 metres, averaging 10 m, but this commonly increases with decreasing complexity of the hyaloclastite flow units. In sections cut southwest-northeast the sediments may be locally absent, leaving the basalt in direct contact with layered hyaloclastite tuff and other members of the hyaloclastite sequence beneath (Sections 2, 3 and 26, Fig. 4), or, in a few places, overlies certain lava flows (e.g. Sections 40 and 42, Fig. 4). The regular columnar joints are smaller in size (diameter 10-20 cm) than in lava of the Older Series, and are usually arranged perpendicular to the lower cooling surface (Figs. 13, 14 and 15). Exceptions from this regularity are described below. Compositionally the basalt is dark grey with a texture varying from aphyric to microporphyric and non-vesicular in the lowest part. Coarse-grained, highly porous varities may occur in central and upper portions. Olivin and plagioclase phenocrysts are present,

b. Irregular jointed basalt/kubbaberg (K): This facies resembles the hackly-jointed "kubbaberg" basalt described from the Older Series. It usually defines the transitional zone (5 to 40 m thick) between the lower massive basalt (C) and the upper hyaloclastite-breccia (HB) division, with relatively sharp, but very irregular contacts (Figs. 13, 14 and 15). Internally the facies represents a compact mass of irregularly oriented, lobate and fanning or rosett-shaped basalt (Fig. 14). All gradations between such compact lobate kubbaberg basalt, with no interstitial hyaloclastite, and lobate basalt fully enclosed in hyaloclastite breccia, have been recognized. At some localities small pockets of hyaloclastite are present in the upper part of the compact basalt. Closely packed lobes are near-circular in outcrop. In cross section they consist of an outer concentrically jointed zone (0.5 m thick), an intermediate zone of massive or partly vesicular basalt (1-2 m) and a central porous and/or hollow zone containing much secondary zeolite and calcite. Contrary to ordinary pillows and lobes fully embedded in hyaloclastite breccia, these lobes lack chilled features, such as glassy crusts on their outer surfaces. Sometimes the elongated basalt lobes, issuing from the lower massive basalt, branch upwards through the kubbaberg

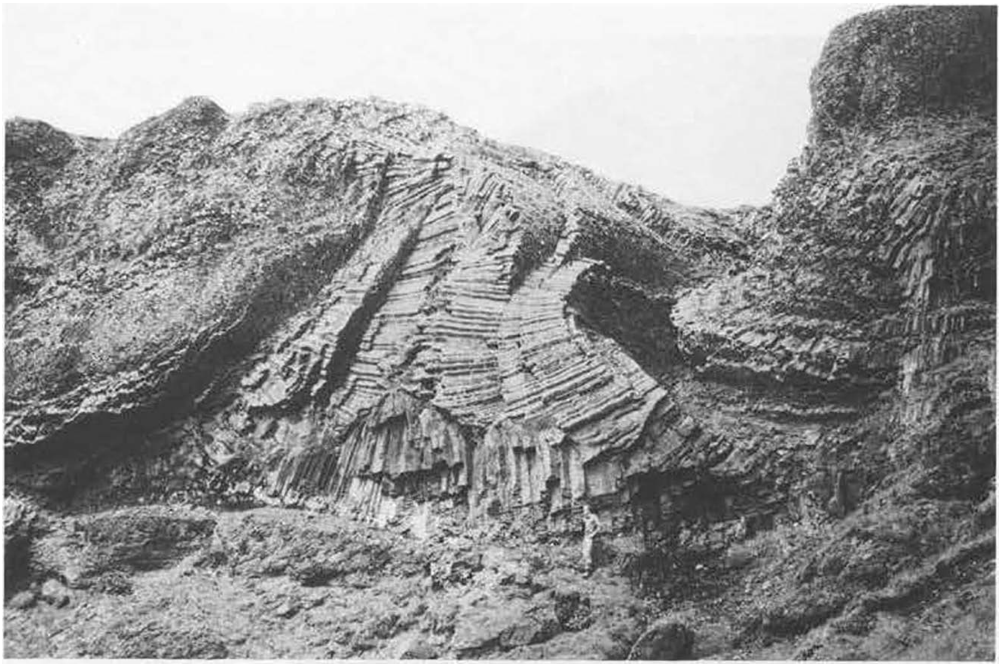


Fig. 13. Irregular basalt slivers at the base of Hy-3C, Seljaland (Section 35), fanning upwards and sideways through hyaloclastite breccia.

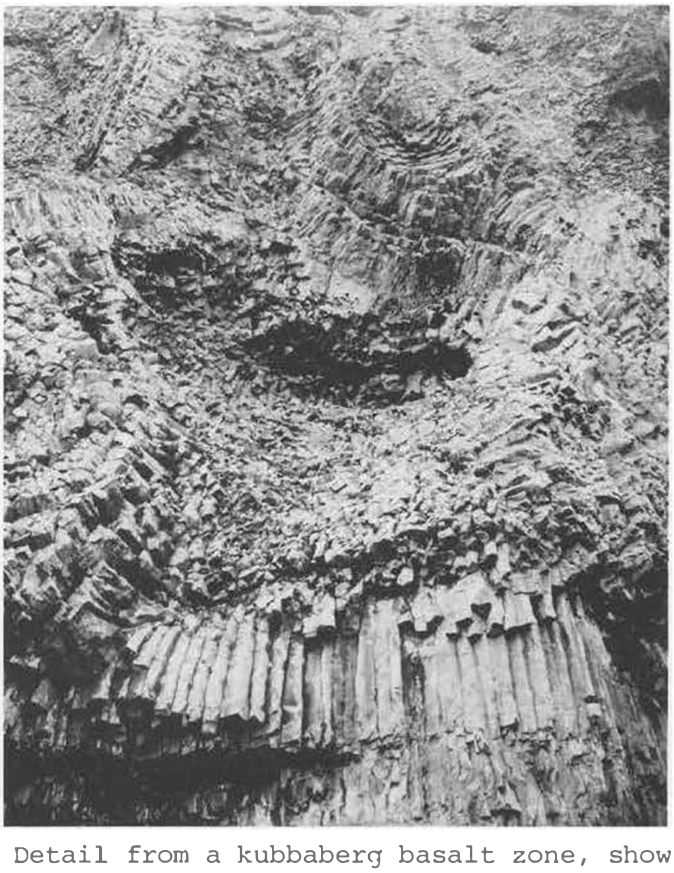


Fig. 14. Detail from a kubbaberg basalt zone, showing compact lobate and fanning/rosett-shaped associations in Hy-4, east of Foss, Section 17.

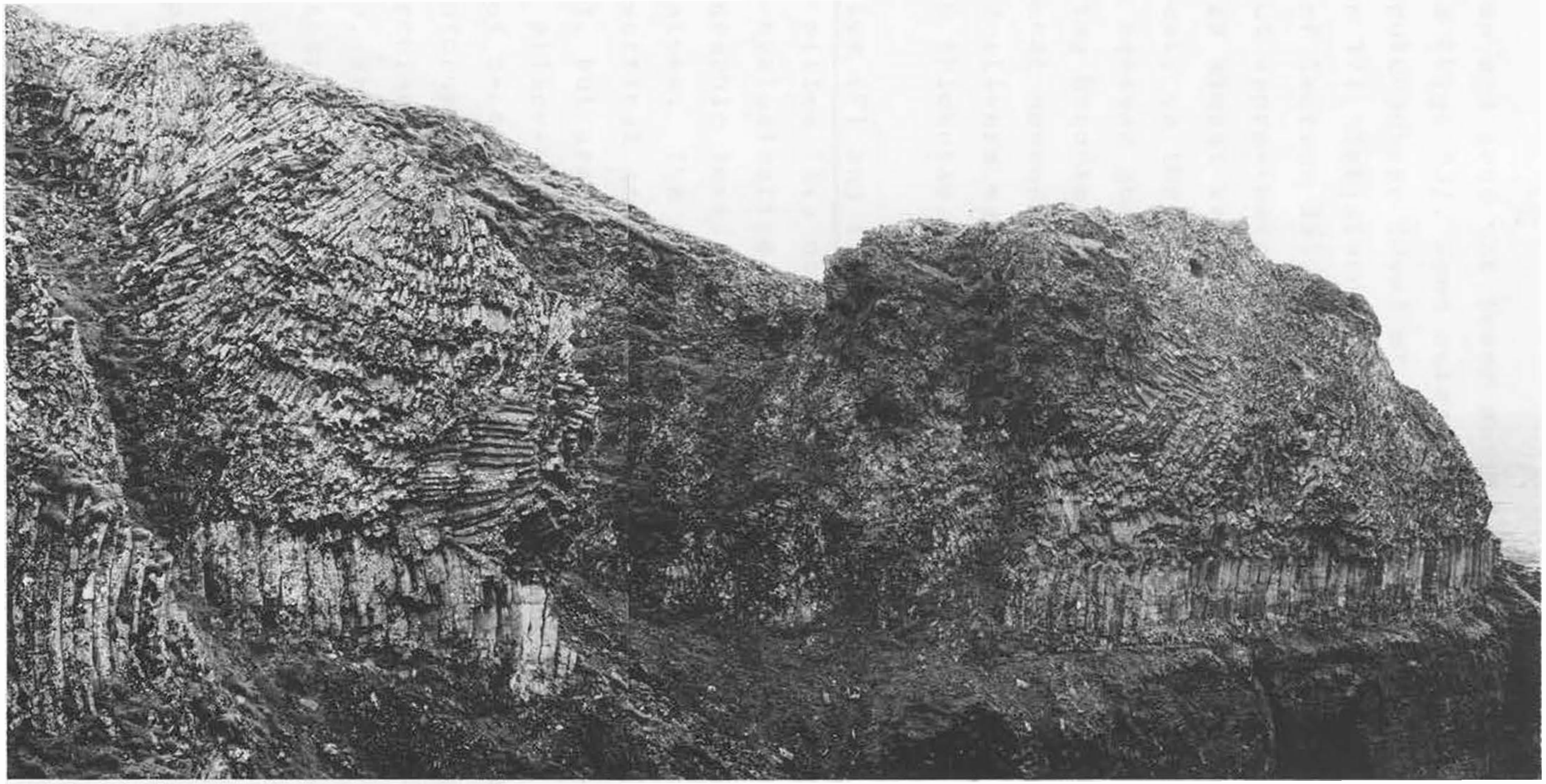


Fig. 15. Basal regular jointed basalt resting sharply on sedimentary diamictite, and being overlain by irregular rosett-shaped kubbaberg facies and hyaloclastite breccia. Locality, Hy-4 at Dalsfjall, Section 34.

division and into the lower portion of the hyaloclastite breccia (Fig. 13). Good outcrops are preserved in the cliff at Fogrutungubryr (Hy-3 of Section 17), Foss (Hy-3 of Section 17), Seljaland (Hy-3C of Section 35) and Dalsfjall (Hy-5 of Section 34). In the section east of Foss (Section 17), cut approximately southwest-northeast, the lobes are arranged almost vertical. In sections cut northwest-southeast, on the other hand, the branching basalt bodies may be sqeezed and bent sideways and upwards through the overlying breccia mass (Fig. 13), indicating some kind of rotational movement during emplacement (see below). Such rotated slivers may exceed 20 m in length and be some 5 to 10 m in thickness.

c. Pillow (P) and isolated pillow basalt (IPB): Close-packed pillow lava occurs sporadically as the basal facies of the hyaloclastite unit, commonly in the lower stratigraphic levels, where the units infill basement depressions. The close-packed nature seldom continues over large vertical and lateral distances (maximum 300 m along strike), but gradually change into isolated pillow basalt and/or pillow-hyaloclastite breccia. A steep wall at Kleifar (Hy-1 of Section 10, Fig. 4), is the best exposed pillow lava outcrop (Noe-Nygaard 1940). Here most pillows appear sub-circular to weakly elongated and bowl-shaped in cross-section, and with diametres from 0.1 to one meter. Small pillows are generally spherical and infill the space between larger pillows. Some larger ones clearly define flow tubes (1-2 m long), as seen from branching interconnections with other pillows. Small entrances and protruberances from the pillow surfaces are also abundant. Some aligned tubes at Kleifar indicate a local flow direction towards east-southeast. Internally, separate pillows display radially arranged contraction cracks, splitting the pillows into pieces. A dark sideromelane crust generally covers the outermost surfaces (0.5 to 1.0 cm), but this is absent were pieces of the crust have been spalled off to form glassy interspace material. This is similar to observations of reentrant pillow crusts described by Dimroth et al. (1978). The pillows are aphyric and microvesicular. Pillow tubes may be hollow in the centre, indicating that their interior were either drained after the crust solidified (Ballard and Moore 1977), or enriched in volatiles (Wells et al. 1979). Pillows displaying concentrically arranged vesicles, as is normally found in Icelandic pillows (Wells et al 1979), occur in several places (for instance Hy-7 of Section 22, Fig. 4), and in isolated pillow basalt overlying the compact pillow lava. The vesicular pillows have scoriaceous cores and pipe vesicles extending radially towards the surface.

The transition from close-packed pillows (P) and/or homogenous basalt (Fig. 16), into isolated pillow basalt (IPB) and/or pillow hyaloclastite breccia (HB), is seen in nearly all outcrops where pillows are present. Usually the change is marked by budding from larger pillows or pillow tubes, to produce isolated pillows supported by more and more interstitial hyaloclastite (up to 20%). Another is fragmentation and/or brecciation of the pillows to produce broken pillow hyaloclastite breccia (Carlisle 1963). This breccia is then made up of angular vitric fragments and disintegrated pillows. The relative amount of interstitial matrix then normally exceeds 20%. The shapes of isolated pillows are much more complex than in close-packed pillow lava. They exhibit amoeboidal to angular shapes and sizes between a few centimetres up to one meter. Imbrication of elongated pillows is commonly observed (Fig. 17). Flow directions estimated from imbricated pillows at Tverarnupur and Heidharsel correspond with measurements of the pillow tubes (e.g. southeastwards).

d. Hyaloclastite breccia (HB): constitutes the bulk material (approximately two thirds) of the Standard Hyaloclastite Unit. Estimated thicknesses are from 10 to 120 m, with an average of 40 m. The facies is according to field apparence and composition, subdivided into four types; 1. Lobate basalt hyaloclastite breccia, 2. Isolated and broken

Fig. 16. Outcrop view of isolated pillows spalled off from an injected basalt body. Locality, Heidharsel, Hy-2A, Section 7.

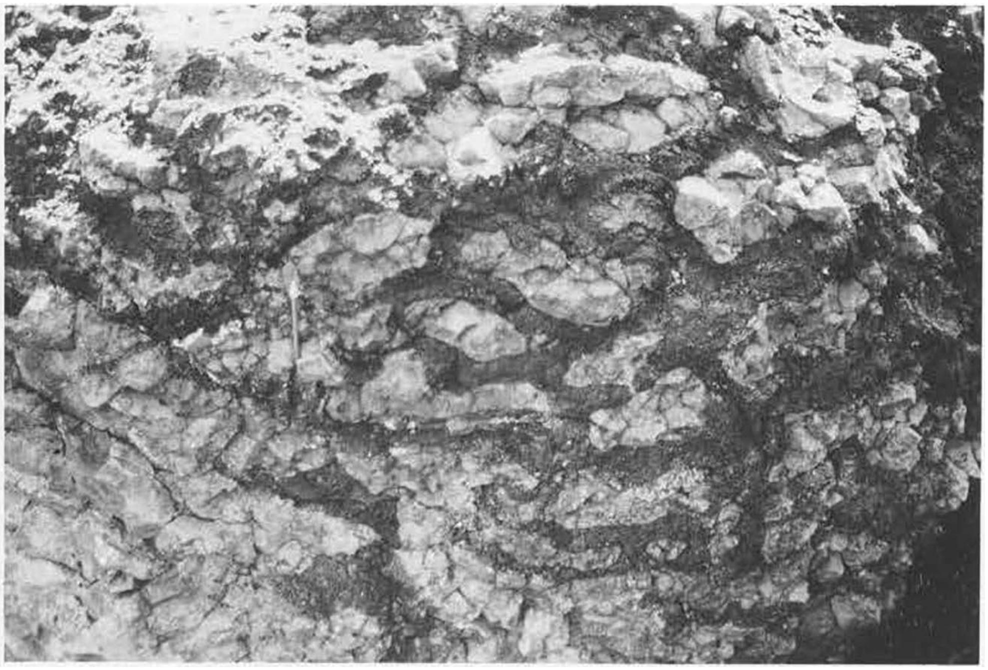


Fig. 17. Imbricated pillows and pillow fragments in Hy-3C at Tverarnupur, Section 18.

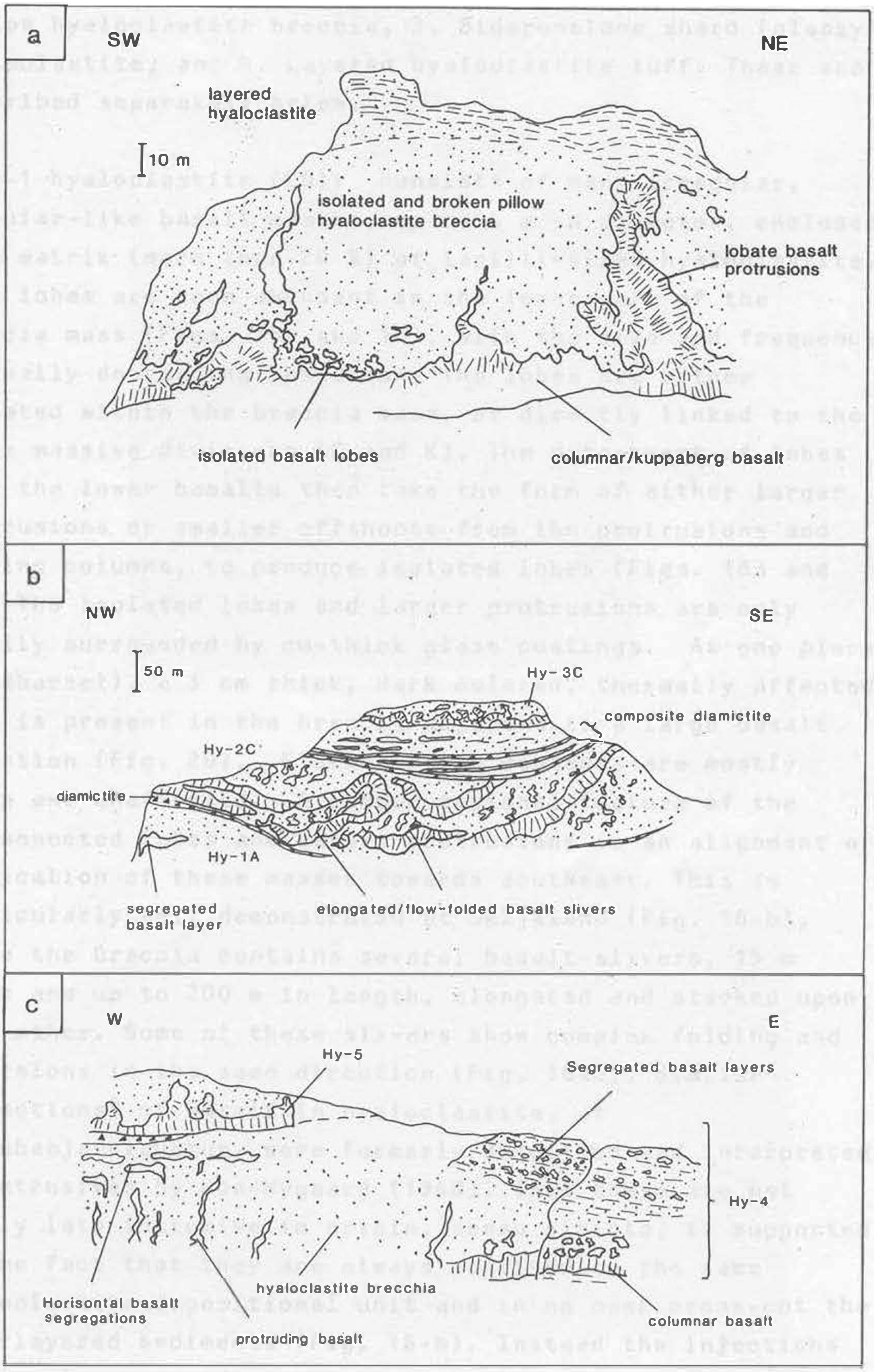


Fig. 18. Drawings after photographs to show large scale features of the hyaloclastite breccia division (LO, IBHB and SSH). See text for explanation.

pillow hyaloclastite breccia, 3. Sideromelane shard (glassy) hyaloclastite, and 4. Layered hyaloclastite tuff. These are described separately below.

Type-1 hyaloclastite (LO): consists of many irregular, globular-like basalt masses, up to 5 m in diameter, enclosed in a matrix (more than 20 %) of lapilli-sized hyaloclastite. Such lobes are more abundant in the lower part of the breccia mass (Figs. 18a and 19), with the size and frequency gradually decreasing upsection. The lobes are either isolated within the breccia mass, or directly linked to the lower massive divisions (C and K). The detachment of lobes from the lower basalts then take the form of either larger protrusions or smaller offshoots from the protrusions and fanning columns, to produce isolated lobes (Figs. 18a and 19). The isolated lobes and larger protrusions are only locally surrounded by cm-thick glass coatings. At one place (Heidharsel), a 5 cm thick, dark colored, thermally affected zone is present in the breccia adjacent to a large basalt injection (Fig. 20). Elsewhere the contacts are mostly sharp and unaffected. A common regional feature of the disconnected lobes and basalt protrusions is an alignment or imbrication of these masses towards southeast. This is particularly well demonstrated at Seljaland (Fig. 18-b), where the breccia contains several basalt-slivers, 15 m thick and up to 200 m in length, elongated and stacked upon each other. Some of these slivers show complex folding and inversions in the same direction (Fig. 18-b). Similar "injections" of basalt in hyaloclastite, at Kirkubaejarklaustur, were formerly described and interpreted as intrusives by Noe-Nygaard (1940). That these are not truely late intrusive in origin, sensu stricto, is supported by the fact that they are always confined to the same hyaloclastite depositional unit and in no case cross-cut the interlayered sediments (Fig. 18-b). Instead the injections nearly always run directly from the lower columnar basalt or kubbaberg members (Fig. 18-a).

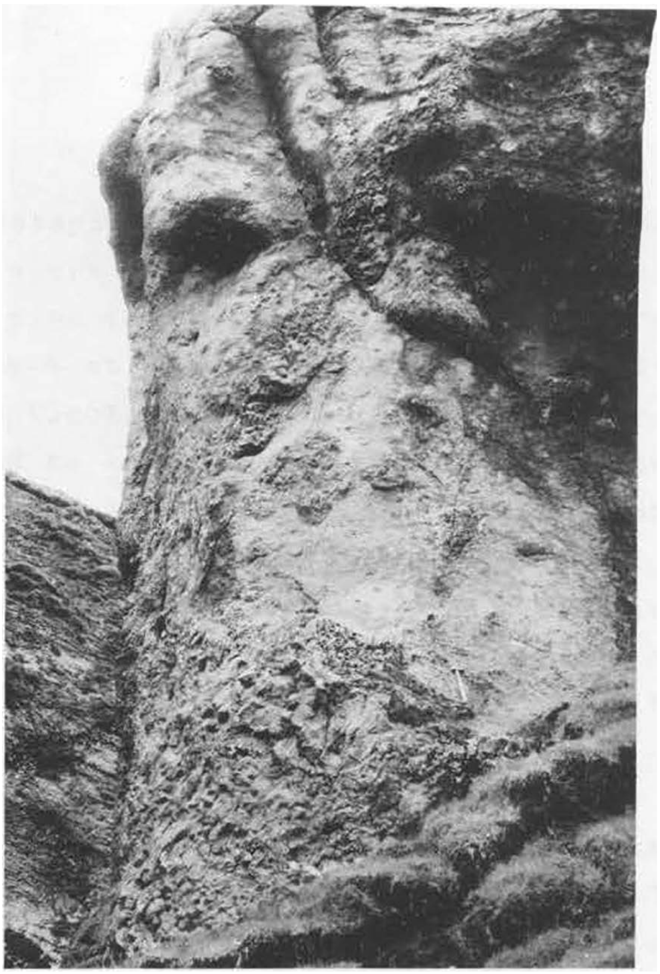


Fig. 19. Isolated basalt lobes spalled off from a vertical basalt protrusion in the lower part of Hy-2 at Kveldsnupur, Section 12.

Fig. 20. Aligned and flow-folded basalt injection in Hy-2A south of Fagrifoss. Note thermal chilling zone (dark grey) of the hyaloclastite surrounding the basalt body.

The basalt masses sometimes arrange themselves into horizontal layers in the upper part of the hyaloclastite breccia division (Fig. 18-c). Excellent outcrops can be studied in Hy-4 at Geirlandshraun (Section 26) and in Hy-1A at Seljaland (Section 35). Such "layers" may be isolated or partly linked to the lower lobate basalt masses. They are sometimes brecciated to form lenticular and/or crescendic layers, up to 10 m thick and 100 m in length, and then comprise coarse vitric material strongly depleted in fines. Similar arrangements have been refered to as segregated layers observed in many pyroclastic flows by Yokohama (1974) and Wilson (1980).

Type-2 hyaloclastite (IBHB): makes up more than 50 % of the total volume of the hyaloclastite flow unit. The facies is completely structurless in most outcrops, with scarcely detectable compositional grading, and consists of a poorly sorted admixture of isolated pillows, disintegrated pillow segments and vitric basalt clasts (Fig. 21). Isolated pillows may reach 0.5 m in diameter, whereas most broken pillows and vitric fragments rarely exceed 20 cm. The matrix (more than 20 %) comprises lapilli-sized dark sideromelane shard hyaloclastite (Fig. 21a), variously palagonized, and with abundant zeolite cement. In the lower part of the Type-2 hyaloclastite the matrix seldom exeeds 40%. The pillow fragments, as shown earlier, are derived both from the larger pillow masses and from the surfaces of protruding lobate basalts, during or after emplacement within the breccia. Similarly, the dark sideromelane shards (up to 3 cm in size) are formed by fragmentation of the glassy crusts of pillow surfaces (Fig. 21a) and lobate basalt inclusions.

Although the breccia appears nearly homogenous in outcrop, there is a weak tendency for upward decrease in the size and abundance of pillows, whereas vitric and broken basalt clasts and the matrix content increase. This change is probably a form of grading.

The vesicularity of individual breccia fragments is a

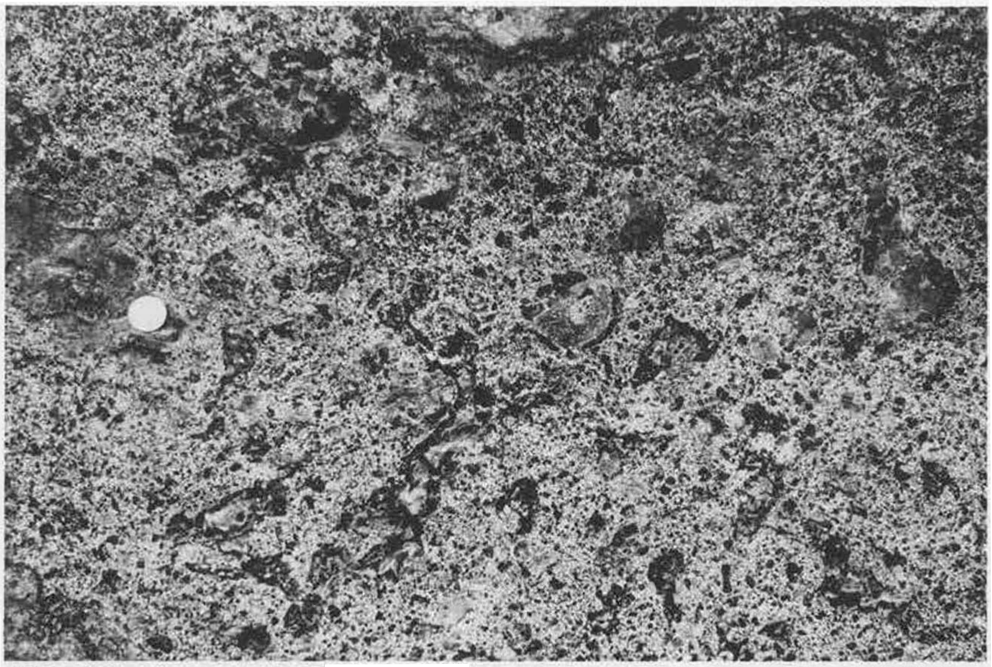


Fig. 21 a. Detailed outcrop view of the isolated and broken pillow hyaloclastite breccia (IBHB). Locality, Kirkubaejarklaustur.

Fig. 21 b. Vitric basalt hyaloclastite breccia with highly palagonized matrix.

function of the total flow thickness and facies association. In thick, highly composite units, the basalt components are dense. Vesicular fragments are more abundant in thin flows and at the top of thick flows. Lithic clasts are only sporadically observed in the upper and lower divisions. Lithics, if present, consist of red or dark grey vesicular basalt (less than 20 cm), considered to be mixed in from the overlying or underlying sediments (see below). Slumped sediment layers sometimes occur in the upper breccia portions. A varity of Type-2 hyaloclastite, consisting essentially (80%) of angular, vitric fragments (Fig. 21b) are common in the upper levels of many flows. The breccia matrix of this particular rock is easily weathered and palagonized (dark yellow color).

In thin section the Type-2 hyaloclastite contains vitric basalt fragments (20-50%) and glass shards (20-40%), tending to be angular with only occassional curved surfaces and relatively few vesicles (Fig. 22). Unaltered glass is only present in the centres of sideromelane shards, whereas the edges and very small grains are invariably altered to gel- and fibro-palagonite (Noe-Nygaard 1940). Platy zeolite (20-50 %) and some calcite constitute the cementing material between the fragmentes. For a more detailed petrographic description of the Type-2 hyaloclastite breccia, the reader is refered to Noe-Nygaard (1940).

Type-3 hyaloclastite (SSH): is a typical light-colored variety of the hyaloclastite breccia, containing abundant dark sideromelane shards (up to 50%) and minor basalt components (15%), and a secondary greyish to white zeolite cement, varying in modal proportion from 40 to 70 %. The light color is due to this intensive zeolitization of the matrix and vesicles of the granules. Chabazite, heulandite and stilbite are the commonest zeolites. Glass shards (0.1-2.0 cm) may be fresh in their centres, or completely palagonized. The glass and vitric components are irregular in shape, with a characteristic rugged morphology bounded by

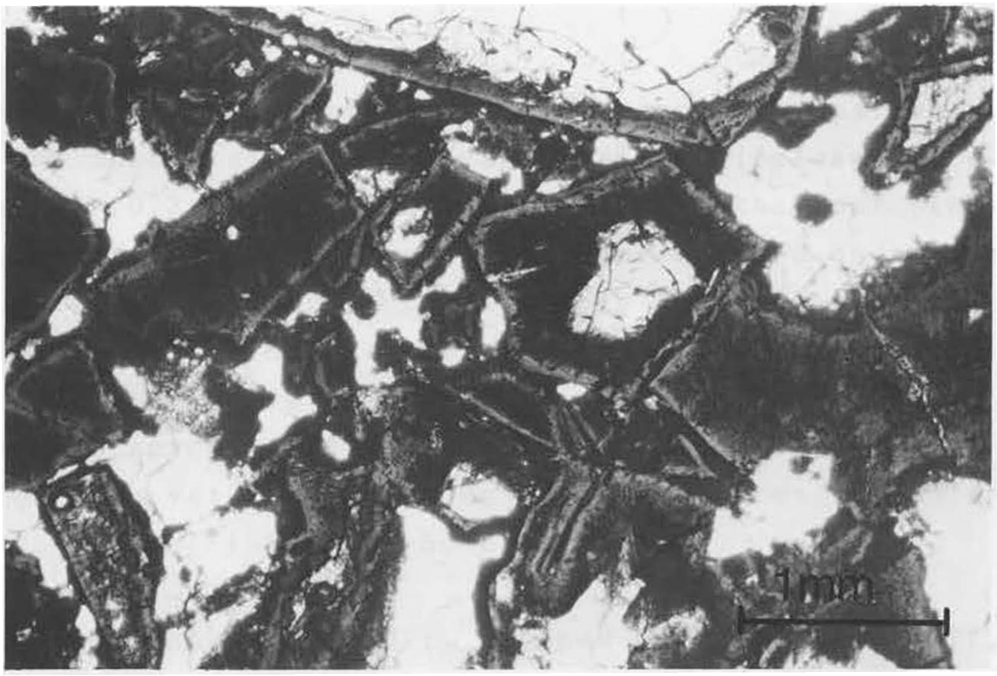


Fig. 22. Microphotograph of Type-2 hyaloclastite showing angular glass shards variously altered to gel and fibro-palagonite. Grains in the upper part have fresh sideromelane in the centres. Matrix between shards consisting of zeolites.

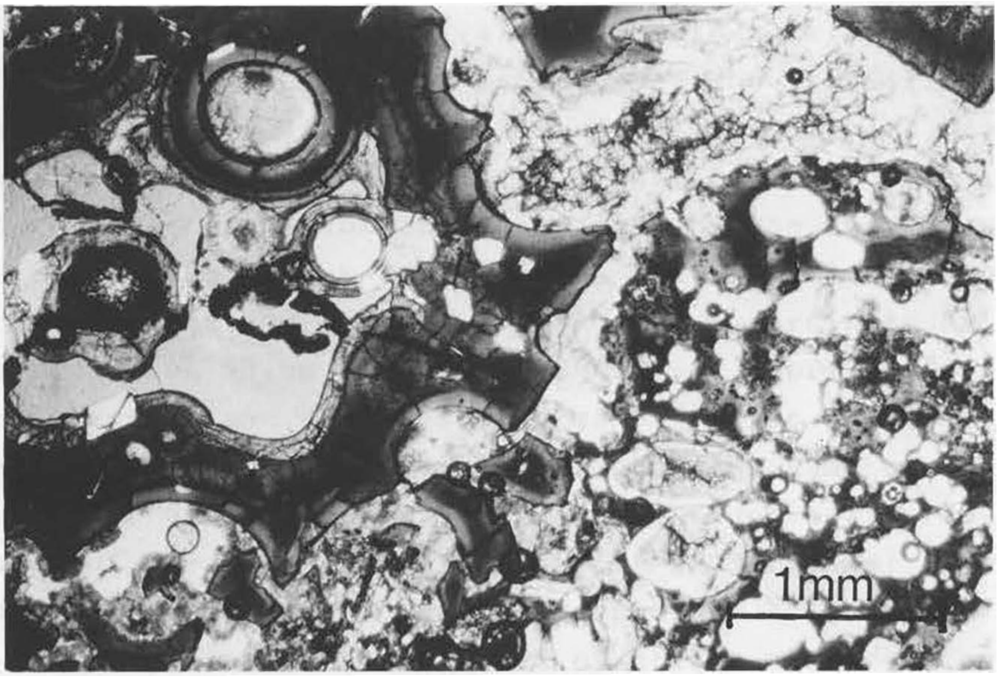


Fig. 23. Microphotograph of glass shards from Type-3 hyaloclastite, displaying concave-convex-shaped grain morphology, high vesicularity and complex secondary palagonitization of the sideromelane.

concave-convex surfaces and contain abundant vesicles (Fig. 23) and spherulites. Some fragmental plagioclase and olivin crystals (0.2 mm) are usually present in the most fine-grained material. The porous and fine-grained character probably made this rock particularly susceptible to alteration. It erodes easily and typically develops a cleavage (secondary compaction), oriented parallell to erosional surfaces. A very common erosional feature of this facies is a vertical split up into structures resembling gas escape pipes (Fisher and Schmincke 1984).

Type-4 hyaloclastite (LH): Layered hyaloclastite tuff, if present, occurs at the top of the Standard Hyaloclastite Unit. The facies locally reach 70 m in thickness (Sections 7 and 44, Figs. 4 and 24), but normally, is no more than 15 m. The bottom contact is gradational into Type-3 or Type-2 hyaloclastite, the upper contact is sharp against sedimentary rocks, or a new hyaloclastite unit. In a few places red basalt fragments are present close to the upper contact, producing a mixed volcaniclastic rock. Commonly the Type-4 hyaloclastite is well-sorted and stratified, with alternating coarse/fine lapilli and ash-tuff layers, ranging from 10 cm to 2.0 m in thickness (Figs. 24, 25 and 26). The texture and composition are controlled by the thickness of individual beds. Thick layers (more than 50 cm) contain vitric fragments up to 3 cm across, and subordinate matrix. Beds less than 50 cm have high content of lapilli-and ash-sized sideromelane granules, and minor vitric particles. Some coarse basaltic fragments of about the same size as in the thicker beds, however, are usually found scattered throughout the finer beds. The matrix (0.3-2.0 mm) is made up of relatively well-rounded sideromelane shards (40-70%) and vitric grains (20%) cemented by either red-brown or yellowish palagonite and an aggregatic zeolite (20-40%). The sideromelane shards are typically coated by dark oxides, and their morphologies are rather complex with rugged, recti-linear outlines, as in Type-3 hyaloclastite (Fig. 23).

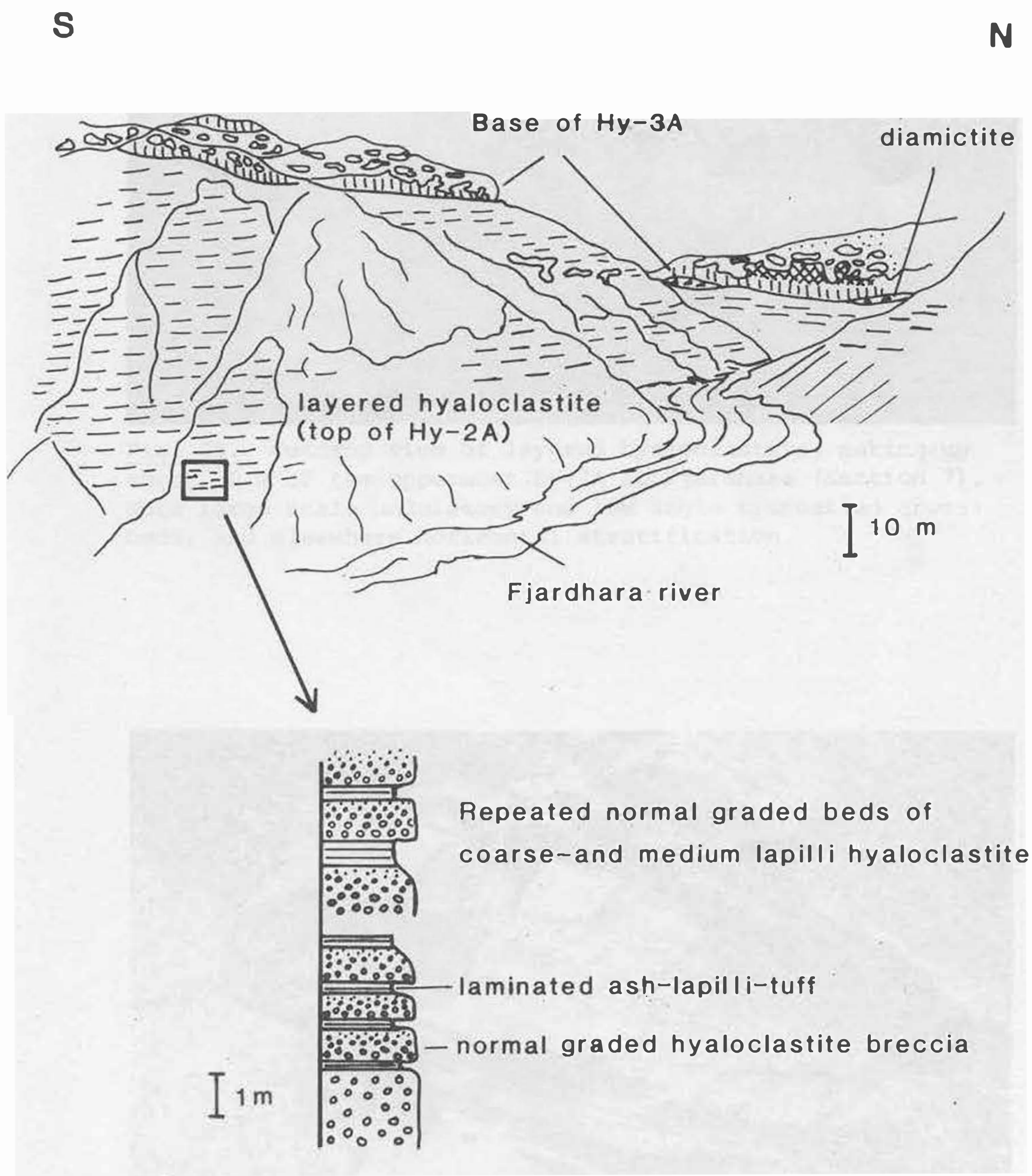


Fig. 24. Sketch of layered hyaloclastite (LH) at the top of Hy-2A at Fjardhara, Section 7. The internal structure and composition of individual layers are shown in detail below.

Fig. 25. Outcrop view of layered hyaloclastite, making up about 50 m of the uppermost Hy-2A at Fjardhara (Section 7). Note large scale undulatory and low angle truncating cross-beds, and elsewhere horizontal stratification.

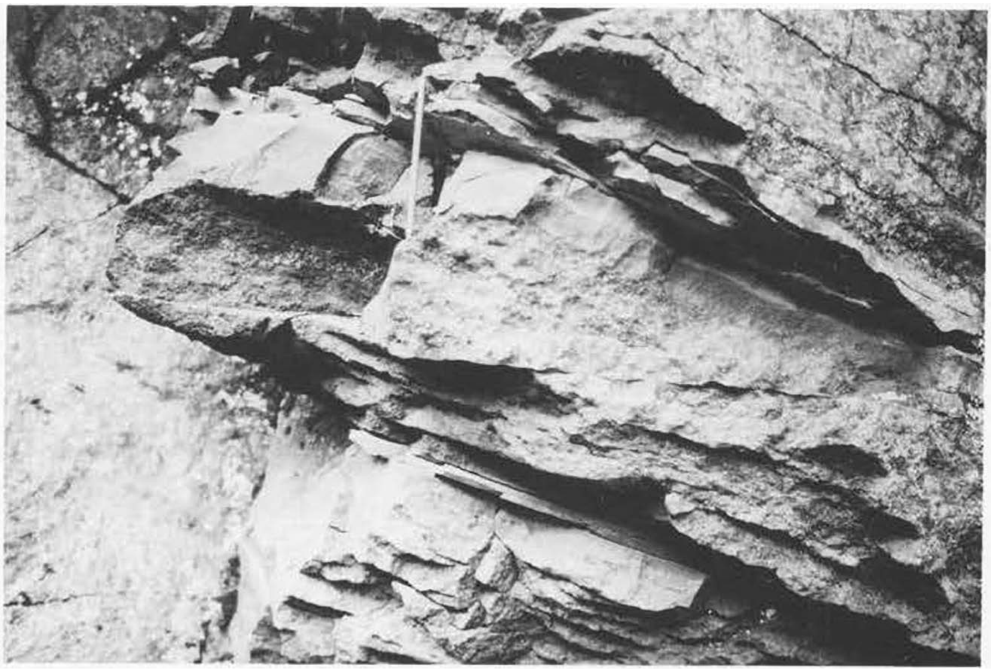


Fig. 26. Details of a normal graded hyaloclastite bed, consisting of a massive normal graded basal member, gradually overlain by thinly laminated ash-lapilli tuff. Locality, Hy-4 south of Kaldbakur.

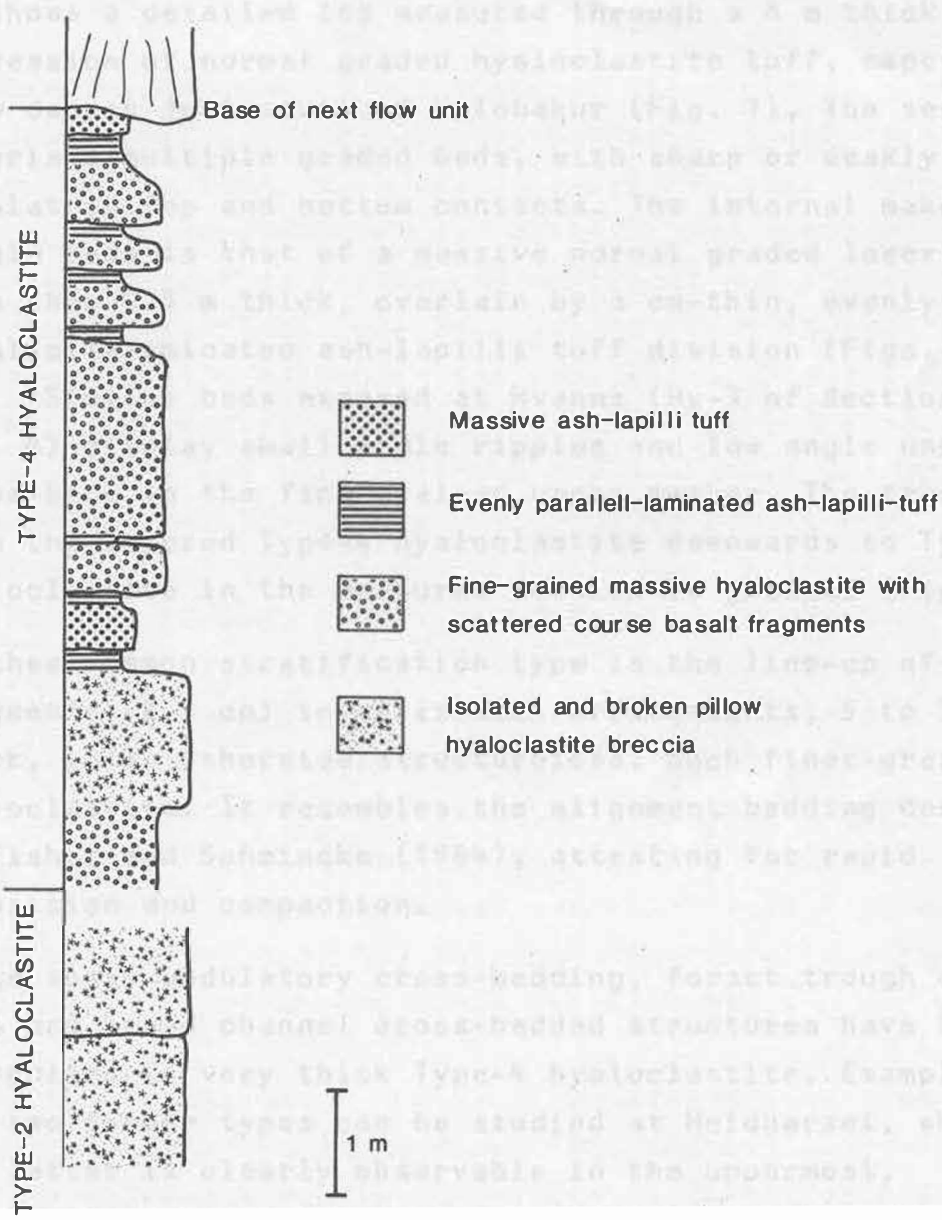


Fig. 27. Vertical log measured through a layered hyaloclastite sequence (Hy-4), exposed south of Kaldbakur.

The arrangement and internal structure of Type-4 hyaloclastite take numerous forms. Commonly, beds appear as multiple, normal graded and/or reverse graded units, 20 cm to 1.5 m thick (Figs. 24, 25 and 26), that are continuous and uniform in thickness over several hundred metres. Fig. 27 shows a detailed log measured through a 8 m thick succession of normal graded hyaloclastite tuff, exposed in a deep canyon just south of Kaldbakur (Fig. 1). The section comprises multiple graded beds, with sharp or weakly undulatory top and bottom contacts. The internal make-up of single beds is that of a massive normal graded lower member, less than 0.5 m thick, overlain by a cm-thin, evenly parallell-laminated ash-lapilli tuff division (Figs. 26 and 27). Similar beds exposed at Hvanna (Hy-3 of Section 38, Fig. 4) display small scale ripples and low angle undulatory cross-beds in the fine-grained upper member. The transition from the layered Type-4 hyaloclastite downwards to Type-2 hyaloclastite in the measured section is gradual (Fig. 27).

Another common stratification type is the line-up of vitric fragments (2-5 cm) in horizontal arrangements, 5 to 30 cm thick, in an otherwise structureless, much finer-grained hyaloclastite. It resembles the alignment bedding described by Fisher and Schmincke (1984), attesting for rapid deposition and compaction.

Large scale undulatory cross-bedding, forset trough cross beds and broad channel cross-bedded structures have been recognized in very thick Type-4 hyaloclastite. Examples of the two former types can be studied at Heidharsel, whereas the latter is clearly observable in the uppermost, southwestern side of Lomagnupur (Hy-6 of Section P44, Fig. 4). Forset beds here display thicknesses from 5 to 60 m and may be up to 200 m in length. The direction of current transport, evidenced from the forset dips, is in all cases towards southeast.

6.2-4. LATERAL FACIES TRANSFORMATIONS:

have been studied in order to obtain regional information about the change in physical conditions during emplacement of the hyaloclastites, and their source area. The tracing of individual units is sometimes difficult, but the occurrence of fairly consistent sediment layers help to fix the stratigraphic position from place to place (Fig. 1).

a. SW-NE variations: The internal structure of most hyaloclastites is fairly consistent in this direction (Fig. 4), but variations, however, do exist. There is, among others, a tendency for simple depositional units to grade into composite flow units, especially at the transition from elevated ridges into paleodepressions. This is usually accompanied by a thickness increase (Figs. 1 and 4). Most commonly the depressions are filled by thick hyaloclastite breccias (HB) and some pillowed and kubbaberg facies, and subordinate amounts of massive columnar basalt (Hy-3A of Section 2, Hy-7 of Sections 3 and 5, Hy-1,2 and 3 of Section 13, Hy-3 of Section 16, Hy-3A of Section 24, Hy-3 of Section 26, and Hy-5 of Section 37). Since the multiple units occur in rapid succession without interbedded sedimentary rocks (for instance Hy-7 of Section 3), they were likely formed during the same eruptive event. On the other hand, the most regular structure association is found in hyaloclastites with nearly constant thickness and relatively flat bases and tops, suggesting that the internal structure is largely controlled by paleotopographic relationships.

b. NW-SE variations: Lateral transformations from composite basalt-enriched flows (K,C and LO) into those containing predominantly brecciated associations (IBHB, SSH and LH) and locally pillow assemblages (P,IPB) are evident in southeasterly directions on a regional scale. This is supported by estimates of the frequency of massive contra brecciated facies in hyaloclastites from different areas, revealing a predominance of massive (C), kubbaberg (K) and lobate basalt breccia (LO) in the northwestern sections, and pillowed (P), brecciated (HB) and layered hyaloclastite (LH)

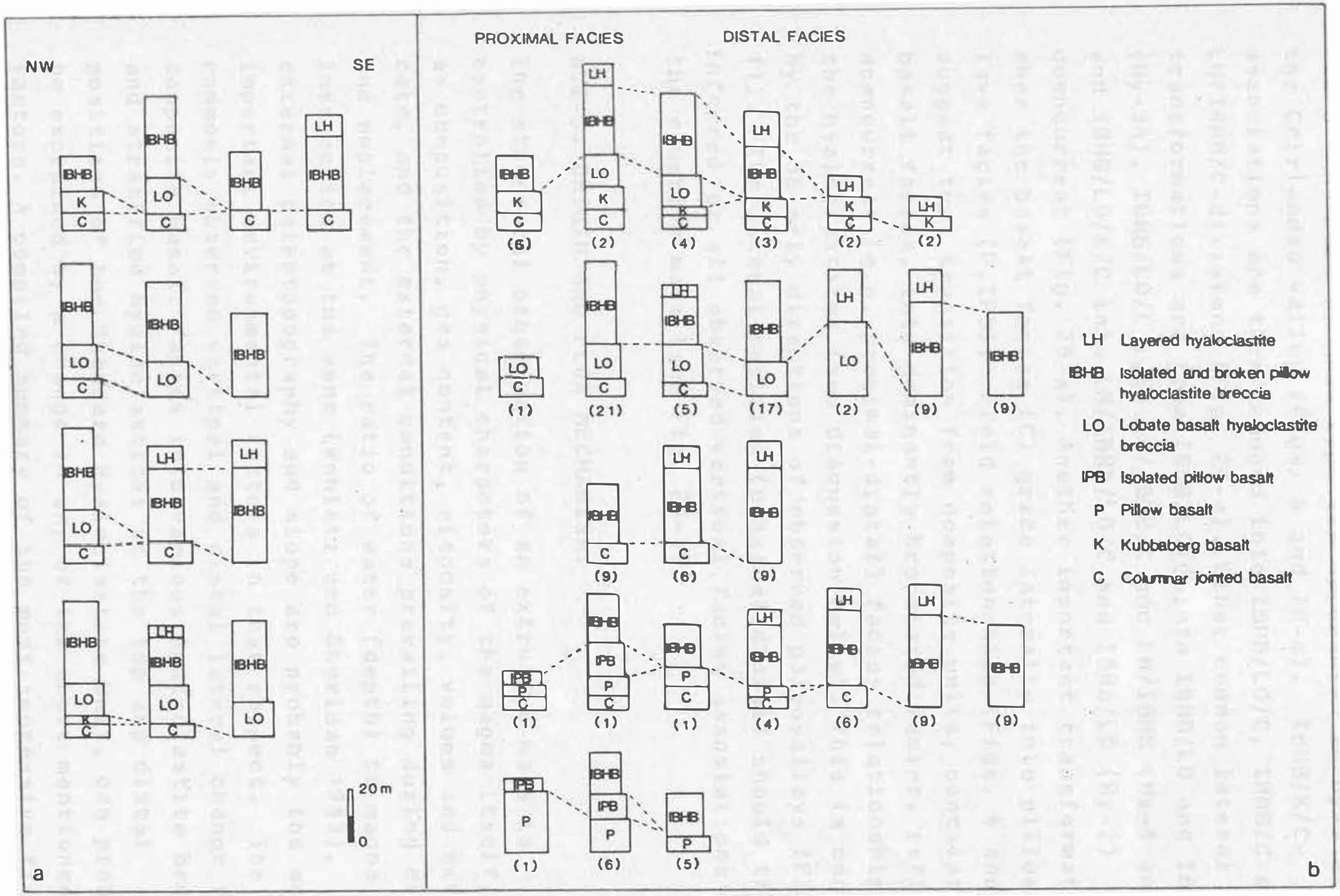


Fig. 28. a. Observed vertical and lateral facies transformations in direction NW-SE within several hyaloclastite flow units. See text for explanation.
b. Suggested proximal-distal facies relationships of the hyaloclastite flows, constructed from all field observations. Number below each column designates how many times the succession has been observed in field outcrop.

in the southeastern sections (Fig. 4). Easily identifiable transformations of this type are documented in Hy-3 along the Geirlandsa valley (Figs. 4 and 28-a). IBHB/K/C-associations are then changed into IBHB/LO/C, IBHB/C and LH/IBHB/C-divisions (Fig. 28-a). Other common lateral transformations are from IBHB/LO/C into IBHB/LO and IBHB (Hy-3A), IBHB/LO/C into LH/IBHB/C and LH/IBHB (Hy-1 and 3A), and IBHB/LO/K/C into LH/IBHB/LO/C and IBHB/LO (Hy-2) downcurrent (Fig. 28-a). Another important transformation is when the basalt facies (C) grade laterally into pillowed lava facies (P,IPB). Field relationships (Figs. 4 and 20-a) suggest that transition from composite units, containing basalt facies, into dominantly brecciated facies, reflects a downcurrent (e.g. proximal-distal) facies relationship of the hyaloclastites (see discussion below). This is confirmed by the SE-erly directions of observed paleovalleys (Fig. 11). The lateral position (proximal-distal) should then be inferred for all observed vertical facies assosiations within the eruptive mass (see Fig. 28-b).

6.2-5. ORIGIN AND FLOW MECHANISM:

The structural organization of an extrusive mass is controlled by physical characters of the magma itself, such as composition, gas content, viscosity, volume and extrusion rate, and the external conditions prevailing during eruption and emplacement. The ratio of water (depth) to magma interaction at the vent (Wohletz and Sheridan 1983), external paleotopography and slope are probably the most important environmental factors in that respect. The commonly observed vertical and distal lateral change from composite basalt facies into various hyaloclastite breccias and stratified hyaloclastites at the top and distal positions of the Standard Hyaloclastite Unit, can probably be explained by a change in any of the above mentioned factors. A compiled summary of the most impressive field and petrographic data described in the previous section and the inferred flow mechanism for the various parts of the Standard

FACIES	COLUMN	FIELD CHARACTERISTICS	FLOW MECHANISM
LH		– multiple normal graded beds – small-and large scale crossbedding – relatively well sorted – alignment bedding	Low-consentration turbulent suspension flow /or turbidite
SSH		– high matrix and shard content – extensive zeolitization light color – local basalt-segregations and fragmental flow layers	Partly fluidized flow
IBHB/ LO		– poorly sorted – internal structurless – matrix more than 20% – large basalt bodies (lobes) resting unsupported in the matrix – aligned and flow-folded basalt-injections – infill topographic depressions	High-consentration dispersive flow (pyroclastic flow/debris flow)
IPB/ P		– isolated pillows with less than 20% matrix – close-packed pillows and pillow tubes – circular and ellipsoide shapes – textural zoned	Low-viscous (sustained) lava flow
K		– irregular jointing – compact lobes/rotated basalt slivers – fanning/rosett-shaped collumns	Turbulent lava flow
C		– regular columns – small size and thickness – extensive	Low-viscous lava flow

Fig. 29. Compiled summary of field characters and inferred flow mechanism for the different facies in the Standard Hyaloclastite Unit.

Hyaloclastite Unit, are shown in Fig. 29, and discussed briefly below.

a. The basal lava facies (C,K,P,): The fact that the basal columnar facies (C) is usually very persistent whereas the others (K,P) have restricted occurrence, and sometimes are confined to paleodepressions in the basement, suggests that the development of basalt facies within the hyaloclastite depositional unit is controlled by the preexisting topography and probably also changing flow rate and water depth. Similar facies variations have been described, among others, in many submarine lavas (Swanson 1972). The development of massive and extensive basalt flows is favoured by high rates of lava discharge and rapid advancement of the flow (Ballard et al. 1979), whereas pillow lava generally forms in deep water by non-explosive interaction of magma and water, at relatively low mechanical energy (Moore 1965, Honnorez and Kirst 1975, Wohletz and Sheridan 1983). With this in mind, the sometimes observed lateral transition from massive into pillowed and kubbaberg basal facies, and furthermore into isolated pillow basalt and/or hyaloclastite breccias, may be explained by the conversion of rapidly moving flows into sustained, more viscous flows, characterized by lower rates of magma discharge and flow motion (Swanson 1972). This change may be due to topographic irregularity, and/or changing processes at the lava front (Cousineau and Dimroth 1982). Under such conditions the already quenched breccia mass may override the lower basalt member to build up thick unstructurated hyaloclastite deposits.

This model is supported by the mapped lateral facies distribution, the very irregular paleotopography, paleoslope orientaton, and prefered occurrence of pillowed and kubbaberg facies in many of the depressions. Another implication is the abundance of multiple facies successions in the paleovalleys, but seldom at higher positions. The composite units found in many depressions were probably formed by inputs of lava supplied to partly inactive channels or channel flow fronts. A similar explanation may account for the repetitions of massive and kubbaberg basalt assosiations, although these are probably formed in a more proximal regime (see later).

Kubbaberg basalt of the Standard Hyaloclastite Unit, unlike those in lavas of the Older Series, is nearly always overlain by Type-1 or 2 hyaloclastite, indicating formation in a fully subaqueous regime (Carlisle 1963). Furthermore, the structural irregularities with fanning lobes, upward and sideways protruding kubbaberg basalt masses and sometimes observed upward rotation of basalt slivers, suggest formation within a lava flow that suffered very high turbulent motion. Similar features have been observed in fluid pahoehoe lava flows (Williams and Mc Birney 1979). This indicates a mode of formation of the kubbaberg which does not fit earlier models of kubbaberg formation. The development of irregular, but still compact kubbaberg structures during the subsequent cooling stage of the lava, implies that the overlying hyaloclastite was initially water-saturated and excerted a pressure high enough to prevent complete quenching of the basalt. Therefore, a steep thermal gradient must have existed across the kubbaberg zone of the Standard Hyaloclastite Unit. On the other hand, a lowered thermal gradient and/or diminishing water pressure, would, possibly, have caused the complete quenching of the basalt to produce hyaloclastite (Carlisle 1963).

b. Brecciated division: The hyaloclastite breccias (LO, IBHB and SSH) consist of poorly sorted material with random oriented fragments (isolated and broken pillows), disconnected lobes and strongly aligned basalt bodies set in a finer grained hyaloclastite matrix. On the one hand, it has been demonstrated that most of the fragmental material is formed by in situ granulation/globulation and spall off from the lower basalt or pillow division, and not by any kind of hydro-explosive eruption mechanism. On the other hand, the complex alignment, large scale folding and very often imbrication of isolated basalt bodies within the unsorted hyaloclastite, suggest formation within a mass that underwent considerable flowage. Since the basalt seldom affects, thermally, the surrounding hyaloclastite, it is assumed that the basalt bodies were emplaced within the quenched hyaloclastite when the material was still hot and fluid, and in movement. The basalt inclusions may have been spalled off from the remaining, massive base during transport, or emplaced in it as a result of later input of magma.

The field and petrographic data furthermore indicate that the main hyaloclastite mass (LO and IBHB) was deposited from a very concentrated suspension of mixed lava fragments and volatiles. The strictly unsorted character with large basalt masses and lobes resting unsupported in the finer-grained hyaloclastite matrix, is typical for high-concentration dispersive flows, such as debris flows (Fisher 1971, Lowe 1976, 1982) and pyroclastic flows (Sparks 1976, Sheridan 1979, Fisher 1979), moving more or less under the influence of gravity. The general absence of internal erosion and/or depositional surfaces implies that the lowermost hyaloclastite breccias (LO/IBHB) were deposited almost continuously from single, very large surges. The large thickness, lateral extensiveness and overall unsupported character imply that the hyaloclastite mass suffered very high internal yield strength, particle concentration and viscosity much higher than that of water. Laminar movement (Johnson 1970, Hampton 1972) of the hyaloclastite breccia division (LO, IBHB), is then favored from the scarcity of grading and sorting.

Sparks (1976) states the importance of laminar flowage in the movement of debris flows and mudflows, but raises objections to this mechanism for pyroclastic flows. This is because pyroclastic flows often contain normal graded lithics and reverse graded pumice clasts, attesting for some kind of sorting mechanism during transport. Wilson (1980) proposes that pyroclastic flows can be typified by fluidization plots, and divides them into three essential types on basis of different degrees of fluidization. Type-one pyroclastic flows of Wilson (1980) are non-or weakly fluidized systems that deposit homogenous material with high concentration of lithics and crystals, and move by near-laminar flowage. The Type-two and three pyroclastic flows are more or less completely fluidized systems that develop normal or inverse grading and localized clast concentration zones due to decreasing flow velocity and particle concentration, and increased turbulent motion (Wilson 1980). In the fluidized flows, particles are supported by an upward stream of gas which is high enough to hinder gravity settling and counterbalance the weight of the flow (Sparks 1978, Sheridan 1979). Fluidization therefore requires a source material with high residual gas content and very large volume.

The influence of fluidization may explain many of the observed internal features of the hyaloclastite breccias (HB) of this study. The upward protruding basalt bodies and lobes are much like injection phenomenon observed in fluidized sediment flows (Middleton and Hampton 1976), and the zones of predominantly vitric material found in upper levels of the sideromelane shard hyaloclastites (SSH), resemble segregation layers found in many fluidized pyroclastic flows (Wilson 1980). It is suggested here that the potential for fluidization of the hyaloclastite breccia division is found both in the massive basalt at the base of the Standard Hyaloclastite Unit, and in external volatiles, e.g. probably the water which quenched the glass during hyaloclastite formation. The latter is a more likely source of volatiles than is degasing alone, since the volume of the upper hyaloclastite breccia in general is many times that of the basal member. However, the presence of vertical erosion structures, resembling degasing pipes (Fisher and Schmincke 1984), could imply that the lower member of the flow, in fact, was quite gas rich, producing an upward stream of gas to the flow surface. A combined effect of the water-and gas-mixing phase was probably to expand the upper brecciated mass, producing a loose, matrix-supported grain fabric (e.g. the Type-3 and 4 hyaloclastite of this work). This hypothesis is supported by the presence of segregated layers in higher parts of the unit. From this it is concluded that the transition from the homogenous, unsorted Type-1 and 2, into Type-3 hyaloclastite may represent the transformation from a laminar dense-phase system (Fisher 1983), resembling the Type-one pyroclastic flow of Wilson (1980), into a more dilute-phase fluidized system, similar to the Type-three pyroclastic flow of Wilson (1980).

The extensive Type-4 hyaloclastite at the top of the Standard Unit, requires further comments at this stage. It is made up of stratified hyaloclastite tuff, involving normal and reverse graded beds, and contains water-induced sedimentary structures. Together, this implies deposition from low-concentration suspension flows (Sparks 1976). Such flows exhibit turbulent motion and are capable of developing graded beds and traction sedimentary currents (Sparks 1976). Turbulence within a moving subaerial pyroclastic flow is largely confined to the flow front (Fisher 1979, Wilson 1980, Wilson and Walker 1982), whereas subaqueous flows may be turbulent (Fiske and Matsuda 1964) at the water-magma surface of the entire flow length, producing extensive turbulent zones. During the initial phases of a subaqueous pyroclastic flow, a thick unstratified lower division is produced from a dense slurry of fragmental debris. In later phases the turbulent cloud may continue to move downslope after the flow has come to rest, resulting in stratified ash, deposited from turbulent suspension (Fiske and Matsuda 1964, Bond 1973, Niem 1977). A similar explanation may apply to the deposition of some of the graded beds in Type-4 hyaloclastite (Niem 1977). On the other hand, the extensive normal graded layers, with uniform thicknesses show internal structural make-ups that are quite similar to distal turbidite sequences (Bouma 1962, Sanders 1965). The somewhat larger grain size of these beds, compared with normal turbity currents, may then be due to the hyaloclastite-flows not carrying large amounts of fine material. This probably reflects the non-explosive mechanism of formation of the hyaloclastites, as noted before, followed by emplacement within somewhat shallower water than usual for turbidity currents.

An alternative interpretation of the normal graded beds is that they were formed by a series of explosive pulses during expulsion of the hyaloclastite mass. Similar inferences were made by Niem (1977) to explain alternate normal bedded ash and pumice tuff overlying the massive division of a subaqueous pyroclastic flow.

A way of testing if the activity switched into an explosive phase towards the end of deposition of the Standard Hyaloclastite Unit, is to study particle shapes of the hyaloclastite. According to Mac Donald (1972), volcanic fragments produced by explosion display many curved surfaces, while fragments resulting from non-explosive granulation contain few vesicles and are generally flat with angular outlines. Figs. 22 and 23 show representative examples of grains from the Type-2 and Type-3/4 hyaloclastite. Fragments of Type-2 hyaloclastite are mostly angular with simple, rectilinear shapes and less vesicular (Fig. 22). Sideromelane shards of Type-3 and 4 hyaloclastite are slightly vesicular, and display complex, concave-convex shaped morphologies with a high percentage of integral gas bubbles (Fig. 23). A binary plot of the parametres N (number of grain corners) against P (planarity %) was used by Honnorez and Kirst (1975) to graphically discriminate the two different kinds of mechanism. This plot (Fig. 30) reliably confirms the texture and grain size observations. There is a gradual transition from Type-2 hyaloclastite, plotting well within the non-explosive field, into Type-3 and Type-4 hyaloclastites, which lie partly above the explosive dividing line (Fig. 30). There is also a tendency for the finer grained fractions (0.2-2 mm) to plot

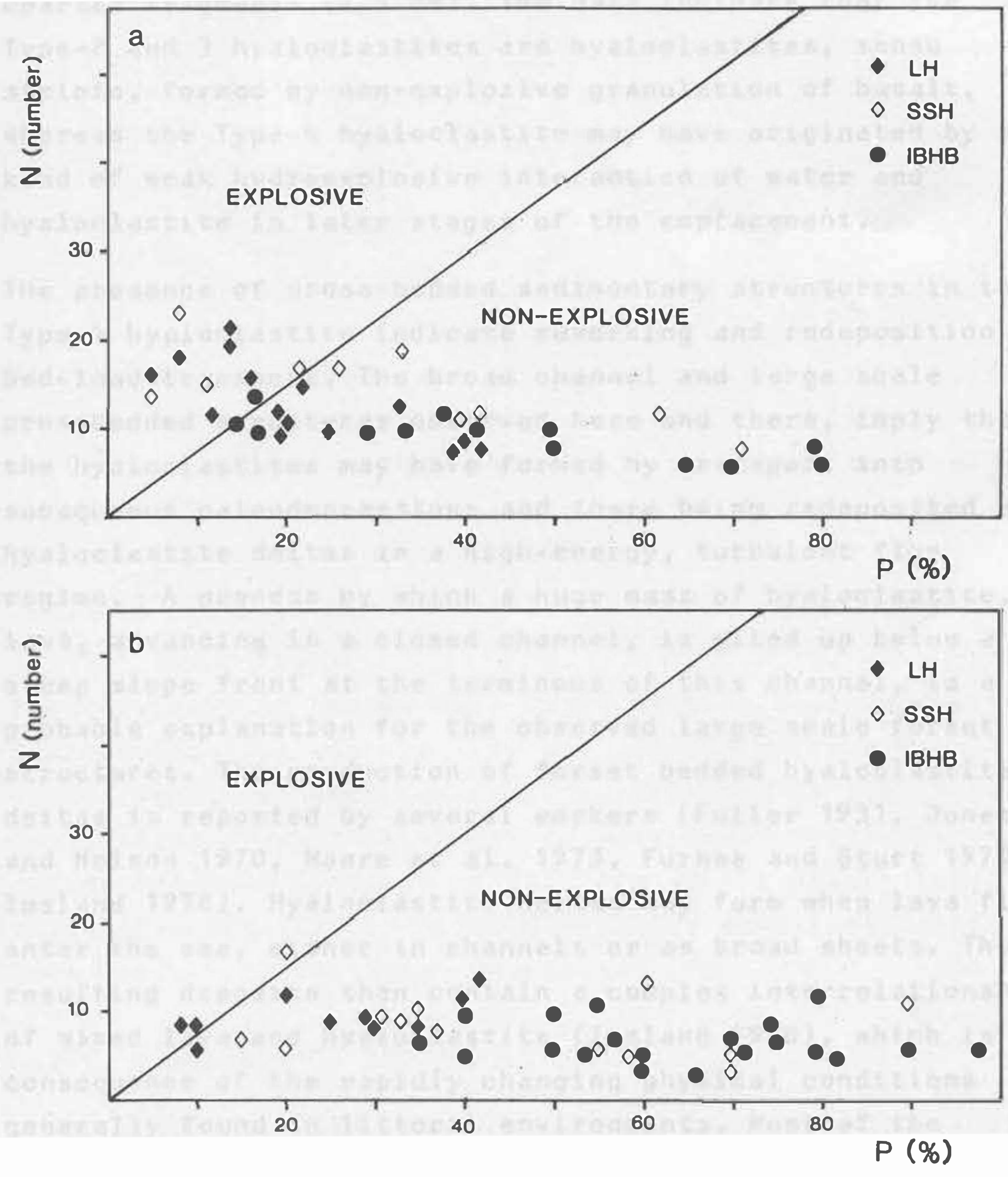


Fig. 30. Grain shapes of isolated and broken pillow hyaloclastite breccia (IBHB), sideromelane shard hyaloclastite (SSH) and layered hyaloclastite (LH), plotted in P-N-diagram (Honnorez and Kirst 1975), discriminating between hyaloclastite formed by non-explosive interaction of basalt and water, and hyalotuff, generated by hyadroexplosive magma-water interaction. a. Grain size range between 0.2 and 2.0 mm, b. Grain size range between 2.0 and 5.0 mm.

more frequently in the explosive field than is the case for coarser fragments (2-5 cm). The data indicate that the Type-2 and 3 hyaloclastites are hyaloclastites, sensu stricto, formed by non-explosive granulation of basalt, whereas the Type-4 hyaloclastite may have originated by some kind of weak hydroexplosive interaction of water and hyaloclastite in later stages of the emplacement.

The presence of cross-bedded sedimentary structures in the Type-4 hyaloclastite indicate reworking and redeposition by bed-load transport. The broad channel and large scale crossbedded structures observed here and there, imply that the hyaloclastites may have formed by transport into subaqueous paleodepressions and there being redeposited as hyaloclastite deltas in a high-energy, turbulent flow regime. A process by which a huge mass of hyaloclastite, or lava, advancing in a closed channel, is piled up below a steep slope front at the terminous of this channel, is a probable explanation for the observed large scale forset structures. The production of forset bedded hyaloclastite deltas is reported by several workers (Fuller 1931, Jones and Nelson 1970, Moore et al. 1973, Furnes and Sturt 1976, Imsland 1978). Hyaloclastite deltas may form when lava flows enter the sea, either in channels or as broad sheets. The resulting deposits then contain a complex interrelationship of mixed lava and hyaloclastite (Imsland 1978), which is a consequence of the rapidly changing physical conditions generally found in littoral environments. Most of the hyaloclastite deltas of this study consist entirely of cross-bedded breccias and are confined to distal parts of the hyaloclastite complexes. This suggests that the original basalt mass was already fragmented when it entered the favorable site of hyaloclastite delta formation.

6.3 SEDIMENTARY ROCKS

Sedimentary rocks of the Older Series make up a very heterogenous assemblage of alternating coarse-and fine lithologies, and some intercalated volcano-sedimentary deposits. Sedimentary rocks interbedded with the horizontal lavas are generally non-fossiliferous, thin, but extensive and uniform in thickness. Some very thick (150 m) conglomerates are, however, present in the lower portion of the Lomagnupur Section (Fig. 5), but these wedge out southwestwards.

At least four stratigraphically distinct sedimentary horizons are present in the Younger Series (Fig. 5). Petrographically these are quite homogenous, but exhibit considerable sedimentological, structural and thickness variability. Despite this variability, most rocks display fairly uniform layers with large lateral extent and a general dip towards southwest (Fig. 1), like the hyaloclastites. The boundaries are generally sharp and irregular, but exhibit no major erosive breaks.

The treatment of sedimentary rocks in this text focuses on sedimentological aspects that are important for making environmental reconstructions. Subdivision of lithofacies composition is made as simple as possible. The lithofacies coding proposed by Eyles et al. (1983), for diamictite sequences, and that of Miall (1978) for sandstones, have been used in classifying the rocks.

6.3-1. CONGLOMERATE AND PEBBLY SANDSTONE:

occur as conformable interbeds in the lower exposed sections of the Older Series (Fig. 5). Figure 31 shows a detailed section measured through the sequence from Lomagnupur. The succession contains discrete beds, 2 to 15 m thick, of conglomerate, stratified pebbly sandstone and some lens-shaped sandstone and mudstone interbeds.

1. Conglomerates; are usually clast-supported and occur either as separate beds (less than 10 m thick), with gradational upper and lower contacts, or as concentrated lenses displaying channel erosive bases (Figs. 31 and 32). The deposits are internally massive and relatively poorly

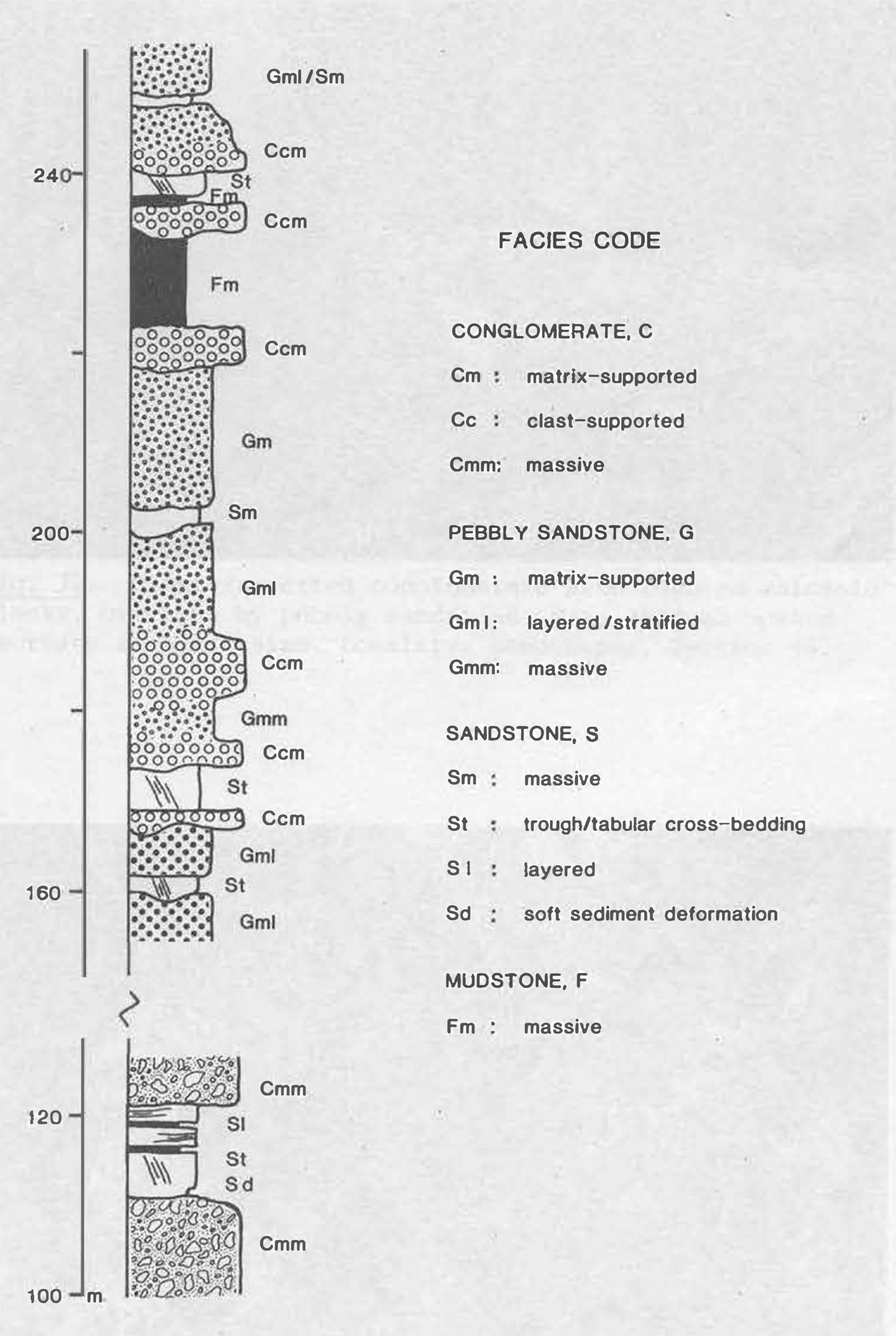


Fig. 31. Section measured through a conglomerate-sandstone succession in the Older Series at Lomagnupur. Characters of the bed contacts are as outlined in Fig. 5.

Fig. 32. Clast-supported conglomerate with rounded volcanic blocks, overlain by pebbly sandstone. Note gradual upward decrease in clast size. Locality, Lomagnupur, Section 44.

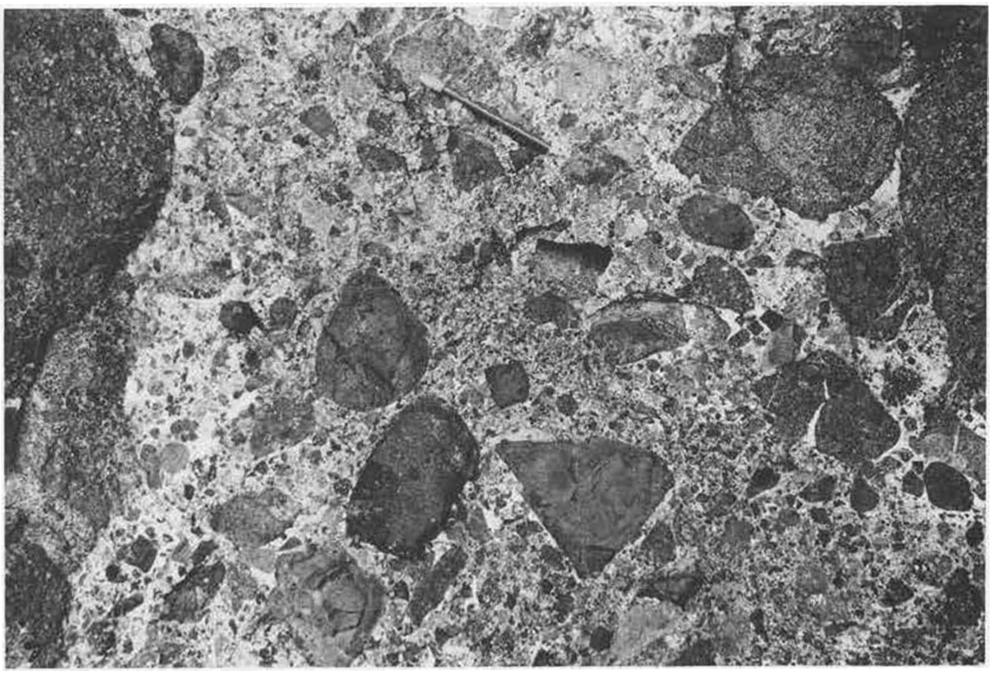


Fig. 33. Matrix-supported, structureless conglomerate from Seljaland (Section 35), making up an integral part of the sedimentary diamictite unit below Hy-3C.

sorted with a polymict clast composition generally exceeding 80% of the total rock volume (Fig. 32). The clasts are significantly well rounded gravel and blocks, 20 cm to one meter in diameter, averaging 40 cm. The polymict clast population includes a variety of rock types; aphyric grey and brown basalt, red amygdal basalt, resedimented conglomerate, sandstone and some mudstone clasts. These are lithologies commonly found elsewhere in the Older Series and in the Tertiary rocks further east. In a few places the channelized conglomerates show normal size grading of clasts, with an upward transition into pebbly sandstones and sandstones (Figs. 31 and 32). Reverse grading is sometimes seen in the flatlying beds.

Peculiar, very coarse matrix-supported conglomerates often occur as integral parts of the sediment layers, both in the Older and Younger Series. Good outcrops can be studied at Nupar (Section 36) and at Seljaland (Section 35, Fig. 4). At Seljaland this conglomerate (below Hy-3C) is 10 m thick and consists of poorly sorted blocky, angular to subrounded volcanic detritus with a maximum clast size of 1.5 m (Fig. 33). The fragments are randomly supported in a fine-grained, highly weathered and zeolitized, brown to red, "earthy" matrix. The amount of clasts approximately equals the matrix. The largest clasts are abundant in the upper levels of the conglomerate, where the rock is gradually overlain by a new lava flow. The lava then gives the impression of being fragmented at its base, with the loosened blocks having sunk down into the breccia, implying rapid deposition upon a still muddy substratum. The matrix-supported conglomerates show restricted lateral extent. The example from Nupar is approximately 50 m in width, grading laterally into finely laminated mudstone and sandstone.

2. Pebbly sandstone: of the Older Series typically occurs in rythmic interbeds (20 cm to 5 m thick) with sandstones and coarser conglomerates, producing an overall stratified succession (Figs. 32 and 34). Pebbly sandstone beds display a somewhat higher proportion of matrix (up to 50%) and

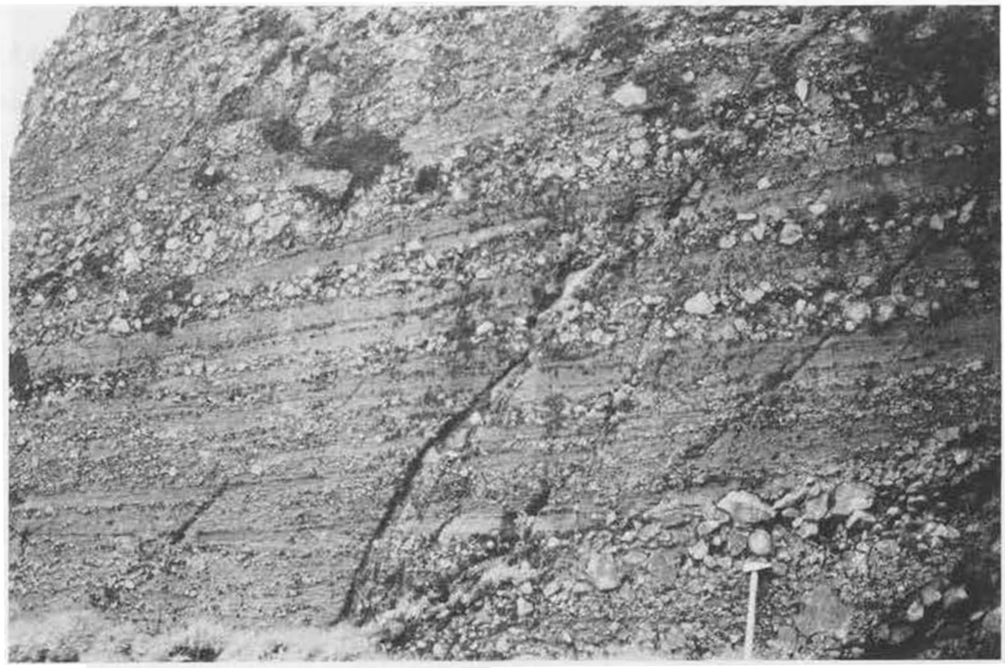


Fig. 34. Pebbly sandstone and sandstone/mudstone interbeds from Lomagnupur (Section 44).

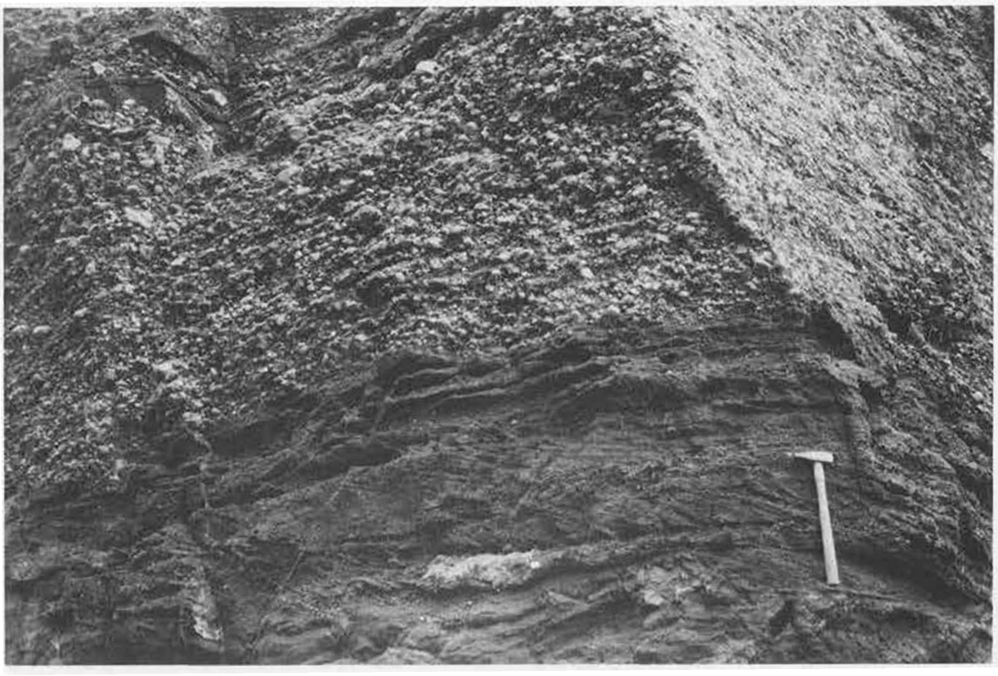


Fig. 35. Detail from a lens-shaped sandstone interval within pebbly sandstone at Lomagnupur, showing small scale tabular cross-beds and flat truncating erosional surfaces.

smaller clast size (5 cm) than the conglomerates. Clast composition is relatively monomict, consisting predominantly of grey aphyric basalt. The sand-sized matrix is made up of altered volcanic debris, and some secondary zeolite. The color is usually brownish. Clasts are well-rounded gravels, mostly randomly arranged. Locally, imbrication of clasts is observed. Tabular-and trough cross beds are abundant in the sandstone intervals (Fig. 35). Such lenses are usually between 20 cm and 4 m thick and with a length of up to 20 m. Cross beds then appear in cosets with several flatly truncating surfaces. A few measured forsets in a lens-shaped sandstone at Lomagnupur and Kalfafell indicate deposition by currents running to the south and southeast, a transport direction quite analogue to that for hyaloclastite deltas of the Younger Series. Large scale cross beds have not been observed in this particular lithology, but are common elsewhere (see below).

6.3-2. PEBBLY MUDSTONE (DIAMICTITE):

Diamictite is used as a descriptive term in this text for consolidated, porly sorted clast-pebble-mud rocks. The name "tillite" generally used for such rocks in Iceland, is avoided since it has a strictly genetic definition, refering to aggregation and deposition by direct glacier ice. Diamictite is the characteristic lithology of sedimentary rocks in the Younger Series, found as intervening layers between the hyaloclastite rocks. Usually the diamictite lithology exhibits very planar and continuous outcrop beds (Fig. 36), but is rather discontinuous where they infill major paleodepressions (Fig. 37), and particularly in southwest-northeast directions. Diamictite layers at higher stratigraphic levels are relatively flatlying. The thickness varies from 2 to 50 m, with an average of 16 m. Individual diamictite horizons usually make up highly composite lithofacies assemblages. Two vertical profiles illustrating the composite nature are depicted in Fig. 38, from Mortunga and Kveldsnupur. The diamictite is a porly sorted, matrix-

Fig. 36. Flat uniform outcrops of two sedimentary diamictite beds, separating Hy-3 and 4 in the cliff at Tverarnupur.

Fig. 37. Large scale paleodepression filled by cross-bedded sedimentary diamictite. View towards south. Locality, north of Husheidi, between Hy-2 and 3, Section 13.

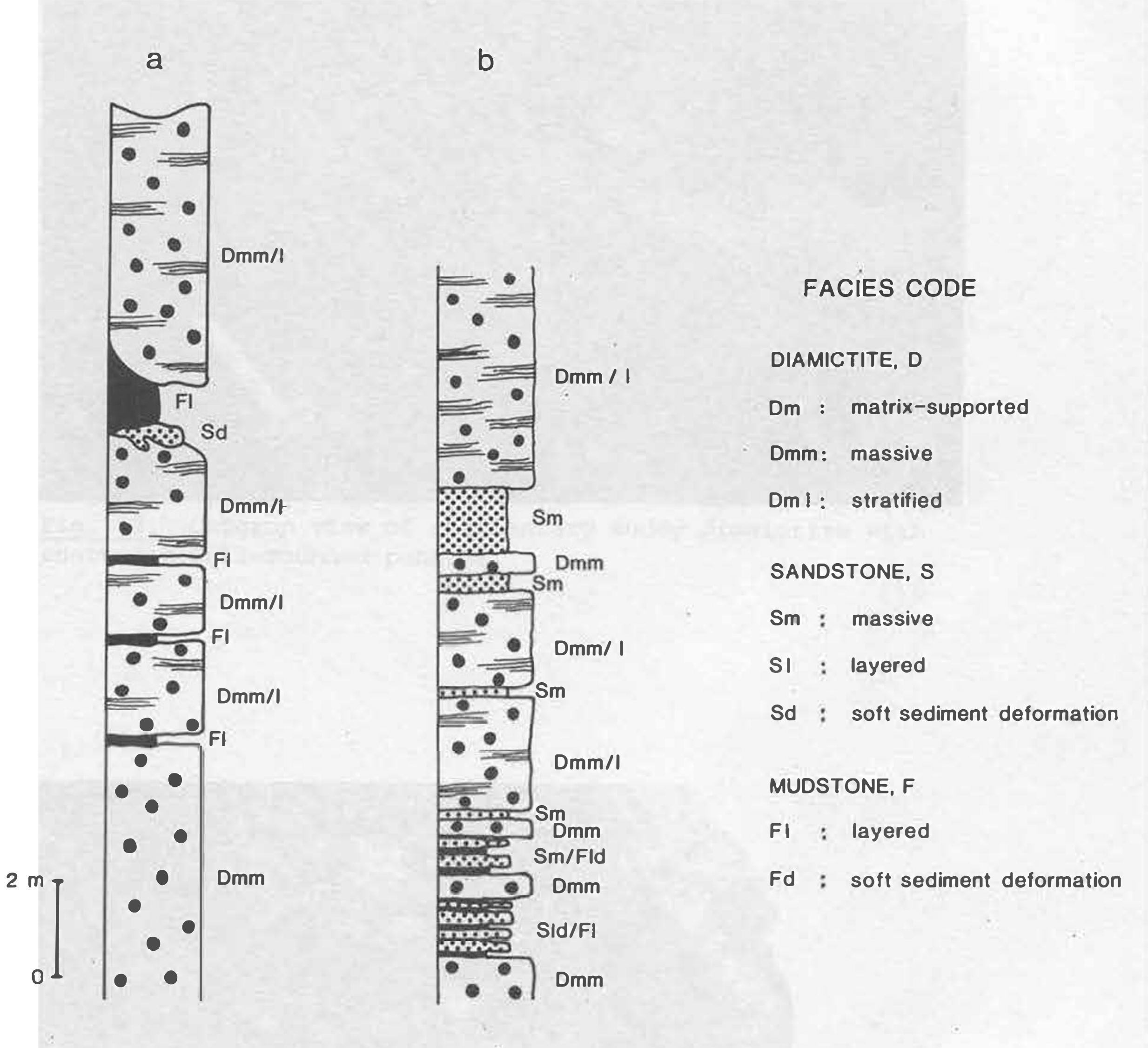


Fig. 38. Vertical lithofacies logs measured through diamictite horizons at, a. Mortunga (below Hy-4, Section 27) and, b. Kveldsnupur (below Hy-2, Section 12). Lithofacies coding after Eyles et al. (1983).

Fig. 39. Outcrop view of sedimentary muddy diamictite with scattered well-rounded pebbles.

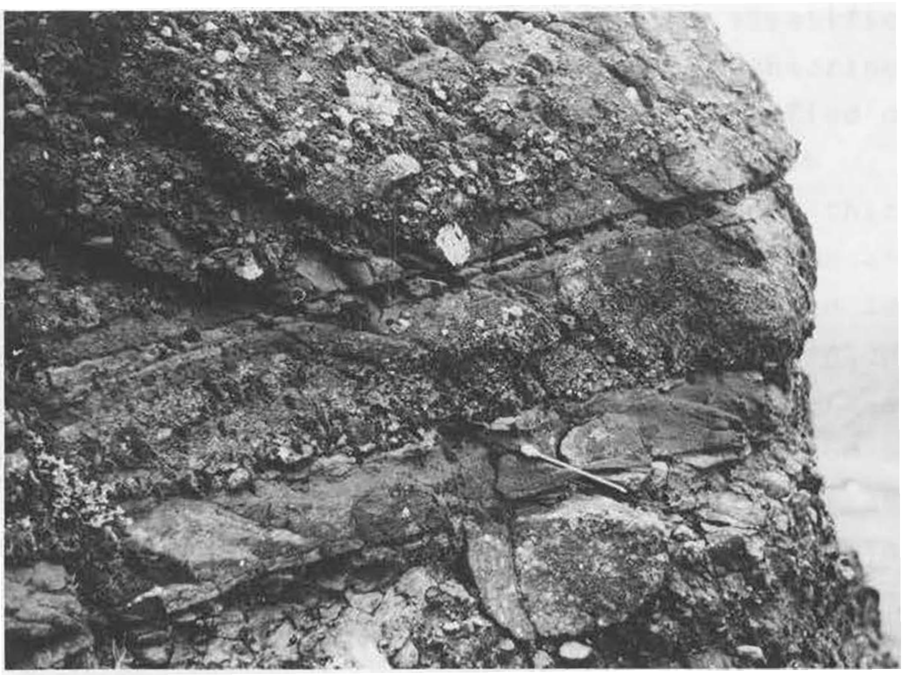


Fig. 40. Sedimentary diamictite with mudstone interbeds at Tverarnupur.

dominated rock that contain scattered pebbles and some blocks, significantly well rounded, enclosed in a dark gray, hardened muddy matrix (Fig. 39). The matrix (60-80 %) varies from clay to coarse sand, but which is usually a fine-grained silty clay. The rock exhibits no grading in either matrix or clast content. The maximum clast size observed is 0.75 m, but commonly range between 8 and 40 cm. Pebbles are monomict, vitric basalt, derived exclusively from the local bedrock. In a few places resedimented sedimentary rocks, pillow fragments and hyaloclastite occur. Striated pebble surfaces are very uncommon if at all excisting. In thin section the matrix consists of a variety of crushed basalt (0.01-0.1 mm), angular sideromelane and some very small corroded olivine and plagioclase-crystals. Zeolite is rarely present, The epiclastic character and the good rounding of pebbles help to distinguish the diamictite from mixed epiclastic-volcanic rocks and hyaloclastites, which appear rather similar in outcrop.

The diamictites are not always completely massive but display some weak internal cleavage-like stratification (Fig. 38), that may be a result of either shearing during deposition, or secondary compaction. Stratified diamictite can thus be regarded as a unit that shows such stratification over more than 20% of the unit thickness. Primary bedding, shown by interbedded sandstone and mudstone (Fig. 40), often with soft-sedimentary deformed lower boundaries (Fig. 38), are also common. On a regional scale there seems to be a lateral change from stratified into massive diamictite southeastwards (Fig. 4). Good sections of bedded diamictite occur just south of Fagrifoss and Kaldbakur (Fig. 1). In the canyon at Fagrifoss, the sedimentary succession is about 30 m thick, and comprises subsequent intervals (1-2 m thick) of different pebble-mud-sand lithologies. Several erosional unconformities, broad channel structures and local flat-truncating cross-stratifications are present.

Very large scale forset cross-bedding is recognized where diamictite infills some of the main paleodepressions. Impressive examples of the latter are seen at Husheidi (Figs. 10 and 37), and just south of Skaftardalur (Fig. 41). In the latter example the forsets are approximately 50 m thick and dip with an angle 50 degrees towards south-southwest. The length of forsets may be some 100 m or more. Internally the forset bedded rock is still a poorly sorted debris without concentration or imbrication of clasts. Soft-sedimentary deformation structures (cm to m scale) are, however, widespread and closely associated with the cross-bedded sedimentary rocks.

The lower contacts of the diamictites vary from sharp, conformable, gradational and/or very irregular. There is, however, a general absence of major erosional surfaces. Incorporation of material from the underlying hyaloclastite is also rarely observed. The passage from hyaloclastite breccia upwards to diamictite is sometimes gradational, and defined by a zone of mixed epiclastic-volcanic rocks (e.g. Section 35, Fig. 4). Interbedded contacts of layered hyaloclastite and diamictite are also seen. Very complex interrelationships occur north of Mortunga, where the contact is substantially convoluted and loaded, and where masses of diamictite have been downfolded into the underlying hyaloclastite. Convolutions of this type are up to 20 m deep and 50 m wide. At the same locality isolated pockets of diamictite, fully enclosed within hyaloclastite, are evident. Some pockets of diamictite then contain highly disturbed and slump-folded sand- and silt intercalations.

6.3-3. MUDSTONES:

occur sporadically between lavas of the Older Series (Fig. 4), and as interbeds within diamictites of the Younger Series (Fig. 40). The most frequent facies is a massive or crudely laminated, sometimes ripple-and small scale cross-bedded silty-clayey rock (Fig. 42), with colors from deep red to brown, yellow and white. The rock is sometimes intensely weathered, others may be hardened with a flinty

Fig. 41. Large scale topset and delta forset crossbedding in diamictite, south of Skaftardalur. The forsets dip towards the viewer. Height of waterfall approximately 30 m.

Fig. 42. Mudstone containing ripples and ripple drift laminations.

consistence, resembling welded ash fall tuff. Welding is, however, disregarded from the presence of structures indicative of current reworking (e.g. ripples and ripple cross-bedding). X-ray diffraction analyses reveal the composition to be a mixture of montmorillonitic clay and zeolite (heulandite, stilbite, chabasite), also found as cement in the adjacent lithologies. Thin layers of red or redbrown clayey or tuffaceous sediments are especially common in association with subaerial lava flows (see previous description). Rythmically laminated, alternating light-and dark-colored clayey rocks, composed of normal and reverse graded units (less than 10 cm) with scattered glass-and basalt clasts, are abundant in the Younger Series. In places where laminated mudstones occur interbedded with diamictite, they may contain rounded "dropstones" of the same lithology as in the diamictite. The laminae are sometimes convoluted and loaded. Fossils have not been recognized so far.

6.3-4. MIXED EPICLASTIC-VOLCANIC ROCKS:

Tuffaceous sandstones and conglomerates are found both as discrete layers (0.2 to 1 m thick) or as transitionary rocks (less than 5 m) between hyaloclastite and diamictite. Some places they are found as isolated lenses and as intercalations within diamictite, and in the upper parts of hyaloclastite breccias and pillow lavas (Noe-Nygaard 1940). A thick succession (40 m) of predominantly thinly bedded tuffaceous sandstone is exposed in the Older Series of the Lomagnupur Section (Fig. 5), gradually overlying sedimentary rocks. The rock is made up of small vitric and crushed pillow fragments and some glass, and an essentially detrital yellow to brown sandy matrix. The detrital matrix and better rounding of the basalt fragments distinguish the mixed epiclastic-volcanic rocks from Type-4 hyaloclastite.

6.3-5. ORIGIN AND ENVIRONMENT OF DEPOSITION:

From the features described above it is obvious that the mechanisms, physical conditions and environment of deposition of the sedimentary rocks must have been quite variable.

a. Older Series: Conserning the sedimentary rocks found at Lomagnupur (Fig. 31), the rapid vertical and lateral change between coarse clast-supported conglomerates, pebbly sandstones and some mudstone, often within the same depositional unit, suggest the importance of sorting processes during sedimentation. Sorting of the material during bed-load transport in running water, seem probable from the presence of closely related channel and cross-bedded structures in sandstones, and the sometimes normal and revers grading found in conglomeratic beds. The rounding of clasts and lack of dispersed fine material in the conglomerates, imply that they are products of long transport in high-energy turbulent regimes, and that they were probably deposited as channel lag. The overlying, and otherwise closely related pebbly sandstones and mudstones (Fig. 31), are likely products of settlement from low-energy suspension and dispersion, operating in the same environment.

A quite different mechanism is considered for the very coarse matrix-supported conglomerates with restricted occurrence, found both in the Older and Younger Series. The extreme chaotic nature, variety and large clast size distribution, matrix-supported character and dearth of sedimentary structures, point to accumulation from flows of relatively high competance, and/or sediment gravity flows (Fisher 1971, Hampton 1972, Middleton and Hampton 1973, 1976). The internal structure and composition are very much like the disorganized conglomerate bed of Walker (1975), suggesting that they were deposited very rapidly on relatively steep slopes.

The sedimentary association from the Older Series at Lomagnupur (Fig. 31) shares many features with alluvial fan/floodplain deposits (Mc Govern and Groat 1971, Bryhni

1978) and/or glacial (sandur) outwash plains (Church 1972), described, among others, from the present day southern coast of Iceland (Ward 1972). Depositional agents on modern alluvial fans and sandur plains are sheetfloods, streamfloods, debris flows and mudflows (Bull 1972). A possible explanation for the succession of stratified sandstones, conglomerates and cross-bedded intervals may, therefore, be that of deposition from repeated sheetfloods and streamfloods. Sheetfloods occur when large amount of water and detritus spread out on alluvial fans/otwash deltas in the form of high-energy sheets (Bull 1972), to deposit extensive normal graded units with well-rounded clasts and little fine-grained material. "Jokulhlaups" are the recent equivalents to sheetfloods on glacial plains in Iceland (Thorarinsson 1969). The coarse, clasts-supported conglomerates may alternatively be interpreted as channel bar deposits of braided streams (streamfloods). The very coarse matrix-supported conglomerates (Fig. 33) could probably have originated from debris flows, deposited in flank gullies or steeply dipping fluvial channels, or be resedimented equivalents of the sheetflood deposits. The absence of fossils is in accordance with the proposed deposition in such a high-energy regime.

The sometimes conspiceous red coloration of mudstones interbedded with the red lavas, confirms a terrestrial origin for sediments of the Older Series. Similar deposits from the Tertiary series in Iceland have been considered as aeolian (Walker 1960, Saemundsson 1979).

b. Younger Series: Sedimentary diamictite characterizing the Younger Series of this study, is traditionary considered to be moraine material (tillite) in Iceland, deposited from grounded or terminal ice. A compact unsorted structure and presence of scouring and striated surfaces on boulders in the rock, are the main arguments for this interpretation (e.g. Walker and Blake 1966, Carswell 1983). Some, but not all of the diamictites described here could possibly represent tillites. However, the fact that most diamictites show relatively thick planar outcrop over areas of 12X25 km without certain major erosive contacts and structural deformation and incorporation of the underlying material, as one normally finds associated with lodgement and melt-out tills (Shaw 1982, Eyles et al. 1983), seem to partly exclude a direct glacial origin. This is confirmed by the occurrence of primary bedding and interbeds of laminated mudstone. Furthermore, the basal contacts into hyaloclastite are always conformable and in many places convoluted and loaded. Eyles et al (1983) regarded similar convolutions of sand layers into diamictite as a result of rapid deposition of sand on a muddy, water-enriched substratum. The substantial loading of diamictite into hyaloclastite observed by us, is interpreted in the same way, by rapid overflow of a mud mass on a still fluid (?) hyaloclastite breccia surface. A substantial amount of water is therefore required to produce such mixtures of mud, sand and pebble, not grounded ice, which normally sorts out the fine material. Many of the resulting diamictites resemble plastic mass flow deposits or debris flows (Middleton and Hampton 1976), induced by gravitational transport.

On the other hand there are some crude evidences suggesting that glacial processes may have been partly active during sedimentation. Examples include the very sparce occurrence of striated surfaces and the U-shaped paleotopography, mainly in the lower stratigraphic levels. Glacial erosion of the Older series prior to the deposition of the Younger Series, is particular evident (Figs. 1 and 10). Since no such major erosional periods are documented within the Younger Series itself, this could probably indicate a retreat of the major ice cap before its deposition.

In a description of Carboniferous glacial diamictites from South Africa, resembling very much the diamictites and diamictite associations of this study, Visser (1983) proposed two depositional models; one in which fluvioglacial sediments built deltas into valleys that had become ice-free after the retreat of ice on land. Subsequent slumping of these oversteepened deltas then produced sediment gravity flows. The other model is that where a partly grounded retreating ice mass deposits till near the ice front and later slumps it down-valley (Dreimanis 1979, Hambrey and Harland 1981, Visser 1983). A combined use of the two models could possibly explain many features of the diamictites treated in this paper. Firstly, the very thick stratified and forset cross-bedded diamictite preserved in paleodepressions implies that the valleys, existing prior to the deposition of the Younger Series, were free of ice, and received large amounts of waterlain debris. Secondly, the heterogenous assemblage of diamictite, mud- and siltstone intercalations, and local delta cross beds often found, indicate that sedimentation occured in repeated successions, not as a continuous process. The separate diamictite "beds" may have formed by subaqueous slumping and sliding of mud on the steep, unstable valley slopes or in delta forsets (Visser 1983). The finer-grained interbeds could have originated by suspension settling from sediment interflows. The unsorted fabric and faintly "sheared" stratification found within the diamictites are further indications of formation from cohesive, water-saturated debris flow-like masses that moved downslope by shearing (laminar flow).

The recognition of scattered rounded pebbles and blocks within both laminated and massive diamictite, suggest that the clasts may have been "rafted" in by floating ice (?) and/or released from glaciofluvial (?) streams and incorporated in lacustrine/shelf mud before the final deposition of diamictite in the subaqueous environment (Fig. 43).

The observed lateral facies change, from dominantly stratified sediments in the northwestern areas of many diamictite units, into massive facies downslope, is in agreement with the present model. Muds may have accumulated in quiet water, for instance a large fresh-water lake, dammed by the glacier (?), with variable ice-rafted, or

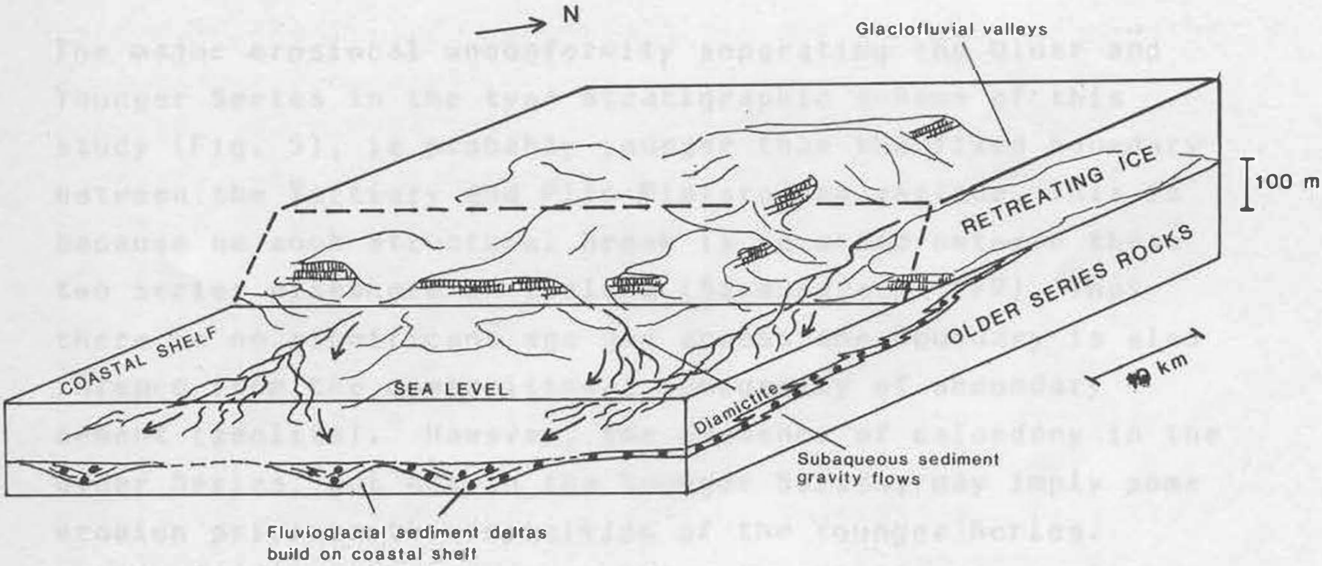


Fig. 43. Plio-pleistocene depositional model for the lowermost diamictite sequences of the Younger Series, in drown valleys along the southeastern coast of Iceland, or alternatively, in very large dammed lakes. The ice-cap responsible for erosion and modification of the previous flat landscape of the Older Series, may have, partly retreated inland.

fluvial transported, components mixed together. Later resedimentation could have been induced by channelward delta progradation and sediment gravity flows, depositing continous diamict material on the subaqueous shelf or in drowned valleys (Fig. 43).

7. DISCUSSION

The major erosional unconformity separating the Older and Younger Series in the type stratigraphic scheme of this study (Fig. 5), is probably younger than the fixed boundary between the Tertiary and Plio-Pleistocene periode. This is because no such structural break is recorded between these two series elsewhere in Iceland (Saemundsson 1979). That there is no significant age gap across the boundary is also inferred from the compositional homogenity of secondary cement (zeolite). However, the presence of calcedony in the Older Series, but not in the Younger Series, may imply some erosion prior to the deposition of the Younger Series. Walker (1960) used the compositional variation of zeolite in lavas to estimate the extent of burial and erosion. Calcedony of tholeiitic lavas corresponds to the chabasite-thompsonite zone of Walker (1960), indicating burial and erosion in the order of 6-700 m. Hence, there must have been a period in the Plio-Pleistocene of this area when the previous quite flat landscape of the Older Series suffered rapid and considerable erosion. The U-shaped paleotopography and the smoothened form of the regional unconformity, indicate that it was a result of glacial erosion.

The stratigraphic context furthermore reveals a marked change in rock products when crossing the unconformity, from dominantly subaerial lavas and terrestrial alluvial/fluvial sedimentary rocks of the Older Series, into hyaloclastites and truely waterlain sediments of the Younger Series (at least in the Sida district). The corresponding change in volcanic and sedimentary rock types then probably reflects a major change in the physical and environmental conditions during deposition, from terrestrial to subaqueous environment. Therefore the environmental conditions prevailing during deposition of the diamictites, may in fact have been the same as those during extrusion and emplacement of the hyaloclastite flows.

The general theory is that the Plio-Pleistocene period in Iceland was characterized by a cold climate when glaciers formed on the island. Hyaloclastites are then considered the products of subglacial volcanism (Noe-Nygaard 1940, van Bemmelen and Rutten 1955, Walker and Blake 1966). What is however, more debatable is whether the whole island, or only parts of it was covered by a single ice cap throughout the period, or if several glaciations and interglaciations may have occurred. Einarsson et al. (1967) recorded ten glacial and interglacial periods at Tjornes in northern Iceland, on basis of recurring tillite and interspersed subaerial lavas. In the study area there is only evidence of one major glacial epoque (e.g. the regional unconformity). From the previous interpretations of the hyaloclastites and diamictites as subaqueous products, this raises the question if glaciers at all existed during formation of the Younger Series in the southern part of Iceland. The principal supporting evidences for suggesting a non-glacial, subaqueous origin of the Younger Series, are; 1. the extensiveness of hyaloclastite and sedimentary rock units, 2. the overall stratigraphic conformity, 3. absence of major erosion surfaces, 4. lack of interbedded subaerial lavas, and 4. the often substantial loaded contacts between sediment and hyaloclastite. The formation of extensive normal graded beds and current induced sedimentary structures and forset cross-bedding, occasionally found in the upper parts of the hyaloclastites, are also difficult to imagine if the area was covered by ice. Furthermore, since the buried paleovalleys in the lower part of the Younger Series are always filled with very thick, truely waterlain sedimentary rocks (diamictite), the valleys could not have been occupied by ice.

There are also certain physical constraints against lateral emplacement of such huge masses of basalt beneath a glacier, with regards to thermal heat effects and triggering mechanisms. Einarsson (1966) calls in question whether the rate of thermal heat transfer from the basalt to the ice is sufficiently high to melt several times their volume of ice, and then to accomodate for the build up of "table mountains". In Walker and Blakes (1966) model, the basalt is believed to have flowed down an ancient ice-covered valley, in which a channel had been created for the passage of water and basalt by the preceeding jokulhlaup (glacial flood). Similarly Nielsen (1937) envisioned englacial tunnels produced by jokulhlaups from the Grimsvotn eruption in 1934. Neither of these considerations, where a glacier is present, can properly explain the transport and deposition of the very extensive (280 km2) sheeted hyaloclastite complexes described here. Therefore we prefer complete subaqueous emplacement of these flows.

An objection against submarine deposition of the sedimentary rocks is the lack of marine fossils, particularly the molluscian fauna so commonly found in Plio-Pleistocene strata of Tjornes (Einarsson et al 1967). No detailed search for fossils were undertaken in this study. It can, therefore, not be excluded that marine and/or fresh water fossils do exist. The character and postulated high-energy regime of deposition of these rocks, however, do not favour fossil accumulation.

If the Younger Series was formed in a fully subaqueous (marine) environment, a significant drop in sea level must have occurred since that time, because the highest altitude of recorded diamictite lies approximately 400 m above the present day sea level. In contrast the maximum Plio-Pleistocene marine limit is known to have been approximately 250 m, corresponding well with the observed erosion of the Older Series. An alternative interpretation for the highest exposed diamictites, may be that they were deposited in an extensive lake with its surface 400 m or more above the present day sea level.

Since there is significant evidence against complete glacial cover during the deposition of the Younger Series, an attractive hypothesis is that the whole series accumulated very rapidly during a late or terminal phase of a major glacial period, which accounted for the erosion of the Older Series. The enormous volcanic productivity following that glaciation may then have been triggered by some kind of glacio-isostatic release in crustal pressure, that gave rise to voluminous extrusions. Extrusion rates in the Plio-Pleistocene of southwestern Iceland is estimated to be some 1000 m/m.y. (Saemundsson 1979). By assuming an average duration of 100000 years for individual glacial and interglacial periods, the build up rate of the Younger Series must have been unique high.

The volcanic source area for the hyaloclastite flows was probably located some 50 km, or more, northwest of the present coastal outcrops. This follows from the mapped areal distribution of the flows, the observed lateral facies changes from mixed basalt-hyaloclastite associations into breccia-dominated facies southeastwards (Fig. 28), the general southeasterly slope of the rocks, the direction of paleovalleys (Fig. 11), and the estimated paleocurrent directions. However, the exact position of the eruptive vent is unknown.

The presumed source area coincides well with the Recent Laki-and Grimsvotn fissure system of the eastern neovolcanic rift zone in Iceland (Fig. 1). Many historical eruptions, originating from this subaerial fissure system, have produced crater rows and widespread lava sheets. The largest and most famous eruption from this system is the Lakagigar lava of 1783. During the Laki eruption approximately 10 cubic km of basaltic lava, covering an area of 370 km2, was produced in 50 days (Thorarinsson 1968). With an active fissure length of 10 km, the average discharge of lava was in the order of 2200 cm3/sec, and the flow rate approximately 5000 m3/sec (Thorarinsson 1968). In comparision, the Plio-Pleistocene hyaloclastite flows of this work, cover areas of up to 300 km2 and maximum volumes of 30 km3. Although the estimated volumes, in fact, do not represent the actual solid magma volume extruded, but rather is an overestimate due to post-eruptive expansion and mixing with water, the volumes of many flows are comparable with that of the Laki 1783 lava. The hyaloclastite flows may therefore be probable subaqueous/subglacial equivalents to the Laki lava, erupted rapidly, from a long subglacial fissure or fissure system.

From the study of lateral facies transformations of the hyaloclastites, it is evident that the eruptions started as effusive outpouring events but changed rapidly into thermally quenched fragmental dispersions, and probable weak hydroclastic eruptions in the final stages of emplacement. This is supported from field observations and by an upward and distal change in the shapes and sizes of involved particle components. This indicates that the water and hyadrostatic pressure conditions during eruption and emplacement of the flows, were quite variable. If the hyaloclastites were generated from a vent in shallow water, and subsequently transported into deeper water, one should expected the whole mass to be more or less completely fragmented. This is not the case. The flows in areas closer to the vent, instead, contain a higher proportion of basalt than hyaloclastite. An explanation may be that an ice cap really existed above the eruptive fissure, yielding a hydrostatic pressure high enough to prevent complete quenching of the basalt. In later phases, however, the mixed lava and hyaloclastite mass may have been transported beyond the ice front, into a shallow subaqueous environment, for instance a wide shelf/fjord, or alternatively a large dammed lake, which caused a release of the hydrostatic pressure, mixing with larger amounts of water, and finally complete brecciation of basalt in the distal flow.

8. HYPOTHETICAL SEQUENCE OF ORIGIN AND EMPLACEMENT OF THE HYALOCLASTITE FLOW UNIT.

The repetition of distinctive hyaloclastite and sedimentary rock types throughout the Younger Series, suggests a repetition of eruptive and sedimentary depositional events. A hypothetical sequence of events leading to deposition of the Standard Hyaloclastite Unit, is illustrated in Fig. 44. This simplified model is based on the observed geological and petrographical features of the resulting deposits, discribed and interpreted in the previous chapters.

1. A voluminous extrusion of basaltic lava from a subglacial/ subaquatic fissure piles up on and around the fissure due to quenching. The very rapid extrusion rate results in a heterogenous mass of solidified fragments and partly isolated products (lobes) of the liquid lava. Gravitational instability causes the whole mass to move downslope, constantly expanding the still liquid lava by fragmentation during flowage. As the flow advances it differentiates into a totally fragmented upper part, which isolates the lower part from direct contact with water, both due to mechanical sealing, but mainly because the hot fragmental unit prevents liquid water from penetrating deep into the mass.

During transportation into channels or valleys, the flow rate may change so that slower, and more viscous basalt facies develop at the base, producing pillow lava and/or differentiated hyaloclastite breccias.

2. During further advancement, the lowermost situated lava members of the flow are increasingly affected by turbulent drag. This may be due to a sudden reduction in the hydrostatic pressure and/or increased water content, or differential movement on the irregular paleotopography. The hyaloclastite-basalt mass moves as a high-concentration, gravity induced suspension flow, much like a subaqueous pyroclastic flow. Density collaps at oversteepened slopes, for instance in subaqueous valleys, resulted in very

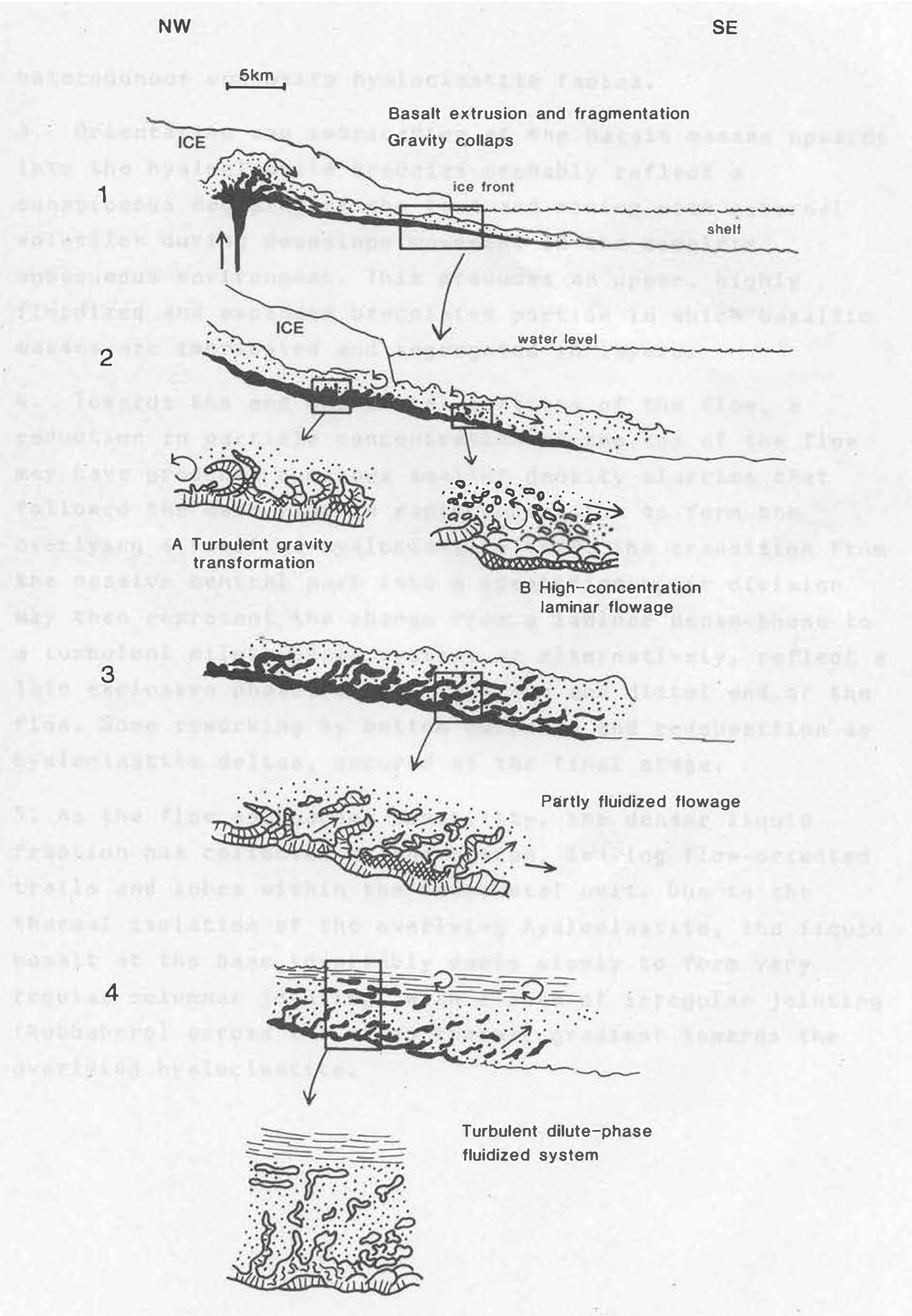


Fig. 44. Eruption, emplacement and flow transformation model for the Standard Hyaloclastite Unit. See text for explanation. The processes operating at the eruptive vent are speculative, because no deposits from this region have been observed.

heterogenous composite hyaloclastite facies.

3. Orientation and imbrication of the basalt masses upwards into the hyaloclastite breccias probably reflect a conspiceous degasing of the lava and mixing with external volatiles during downslope movement in the complete subaqueous environment. This produces an upper, highly fluidized and expanded brecciated portion in which basaltic masses are imbricated and segregated in layers.

4. Towards the end and distal portions of the flow, a reduction in particle concentration at the top of the flow may have produced numerous smaller density slurries that followed the main flow in rapid succession to form the overlying stratified hyaloclastite tuff. The transition from the massive central part into a stratified upper division may then represent the change from a laminar dense-phase to a turbulent dilute-phase system, or alternatively, reflect a late explosive phase, at the very top and distal end of the flow. Some reworking by bottom currents and redeposition as hyaloclastite deltas, occured at the final stage.

5. As the flow approaches immobility, the denser liquid fraction has collected at the bottom, leaving flow-oriented trails and lobes within the fragmental unit. Due to the thermal isolation of the overlying hyaloclastite, the liquid basalt at the base invariably cools slowly to form very regular columnar jointing, with a zone of irregular jointing (kubbaberg) across the steep thermal gradient towards the overlying hyaloclastite.

ACKNOWLEDGEMENT

This project was proposed by G. E. Sigvaldason for my one year fellowship at the Nordic Volcanological Institute, supported by the Nordic Council. G. E. Sigvaldason is acknowledged for introducing me to the field and some of the main problems. The field work was financed by the Nordic Volcanological Institute. My wife Kirsten assisted during field work and is acknowledged for drawing many of the figures. G. E. Sigvaldason critically read the first manuscript and suggested improvements. The other collegues at the institute are thanked for helpful discussions and various practical advice.

REFERENCES

Allen, C.C., 1980: Icelandic subglacial volcanics: Thermal and physical studies. Journ. Geol., 88, p.108-117.

Ballard, R.D., and Moore, J.G., 1977: Photographic atlas of the Mid-Atlantic Ridge rift valley: New York, Springer-Verlag, 114p.

Ballard, R.D., Holcomb, R.D., and van Andel, T.H., 1979: The Galapagos rift at 86 W: 3. Sheet flows, collapse pits and lava lakes of the rift valley. J. Geophys. Res., 84, p.5407-5422.

Bond, G.C., 1973: A late Paleozoic volcanic arc in the eastern Alaska Range, Alaska. Journ. Geol., 81, p.557-575.

Bouma, A.H., 1962: Sedimentology of some flysch deposits; A graphic approach to facies interpretation. Elsevier, Amsterdam., p.1-168.

Bryhni, I., 1978: Flood deposits in the Hornelen Basin, West Norway (Old Red Sandstone). Norsk geol tidsskr., 58, p.273-300.

Bull, W.B., 1972: Recognition of alluvial-fan deposits in the stratigraphic record. In: Rigby, J.K. & Hamblin, W.K. (eds.). Recognition of ancient Sedimentary Environments. Spec. Publ. Soc. Econ. Paleont. Miner. Tulsa 16., p.63-83.

Carlisle, D., 1963: Pillow breccias and their aquagene tuffs, Quadra Island, British Columbia. Journ. Geol., 71, p.48-71.

Carswell, D.A., 1983: The volcanic rocks of the Solheimajokull area, southern Iceland. Jokull 33, p.61-71.

Church, M., 1972: Baffin Island sandurs -a study of arctic fluvial processes. Geol. Surv. Can. Bull., 216, 208p.

Cousineau, P., and Dimroth, E., 1982: Interpretation of the relations between massive, pillowed and brecciated facies in an archean submarine andesite volcano -Amulet Andesite, Rouyn-Noranda, Canada. Journ. Volc. Geoth. Res., 13, p.83-102.

De Raaf, J.F.M., Reading, H.G., and Walker, R.G., 1964: Cyclic sedimentation in the Lower Westphalian of No Devon, England. Sedimenology 4, p.1-52.

De Rosen-Spence, A.F., Provost, G., Dimroth, E., Gochnauer, K., and Owen, V., 1980: Archean subaqueous flows, Rouyn-Noranda, Quebec, Canada, and their Quaternary equivalents. Precambrian Res., 12, p.43-77.

Dimroth, E., 1977: Archean autoclastic volcanic rocks. Rouyn-Noranda area, Quebec: Classification, diagnosis and interpretation. In: Report of Activities, Part A. Geol. Surv. of Canada, Paper 77-1A, p.513-522.

Dimroth, E., Cousineau, P., Leduc, M., and Sanschagrin, Y., 1978: Structure and organization of Archean subaqueous basalt flows, Rouyn-Noranda area, Quebec, Canada. Can. J. Earth Sci., 15, p.902-918.

Dreimanis, A., 1979: The problems of waterlain tills. In: Moraines and Varves (Ed. by C. Schluchter), p.167-179. A.A. Balkema, Rotterdam.

Einarsson, Th., Hopkins, D.M., and Doell, R.R., (1967): The stratigraphy of Tjornes, Northern Iceland and the history of the Bering Land Bridge. In: Hopkins, D.M. (ed.): The Bering Land Bridge: 312-325. Stanford Univ. Press.

Einarsson, Tr., 1946: Origin of the basic tuffs of Iceland. Acta Nat. Islandica, I, p.1-75.

Einarsson, Tr., 1962: Upper Tertiary and Pleistocene rocks in Iceland: Visindafelag Islendinga, 36, p.1-196.

Einarsson, Tr., 1966: Physical aspects of subglacial eruptions. Jokull, 16, p.167-174.

Eyles, N., Eyles, C.H., and Miall, A.D., 1983: Lithofacies types and vertical profile models; an alternative approach to the description and environmental interpretation of glacial diamict and diamictite sequences. Sedimentology, 30, p.393-410.

Fisher, R.V., 1961: Proposed classification of volcaniclastic sediments and rocks. Geol. Soc. Amer. Bull., 72, p.1409-1414.

Fisher, R.V., 1971: Features of coarse-grained, high-consentration fluids and their deposits. Journ. Sed. Petrol., 41, p.916-927.

Fisher, R.V., 1979: Models for pyroclastic surges and pyroclastic flows. Journ. Volc. Geoth. Res., 6, p.305-318.

Fisher, R.V., 1983: Flow transformation in sediment gravity flows. Geology, 11, p.273-274.

Fisher, R.V., and Schmincke, H.-U., 1984: Pyroclastic rocks. Springer-Verlag, Berlin, Heidelberg, New York, Tokyo, 472 p.

Fiske, R.S., and Matsuda, T., 1964: Submarine equivalents of ash flows in the Tokiwa Formation, Japan. Amer. J. Sci., 262, p.76-106.

Fridleifsson, I.B., 1977: Distribution of large basaltic intrusions in the Icelandic crust and the nature of the layer 2 - layer 3 boundary. Geol. Soc. Am. Bull., 88, p.1689-1693.

Fuller, R.E., 1931: The aqueous chilling of basaltic lava on the Colombia River Plateau. Amer. J. Sci., 21, p.281-300.

Furnes, H. and Sturt, B.A., 1976: Beach/shallow marine hyaloclastite deposits and their geological significance. An example from Gran Canaria. Journ. Geol., 84, p.439-453.

Furnes, H., and Fridleifsson, 1974: Tidal effects on the formation of pillow lava/hyaloclastite deltas. Geology, 2, p.381-384.

Furnes, H., Fridleifsson, I.B., and Atkins, F.B., 1980: Subglacial volcanoes - on the formation of acid hyaloclastites. Geotherm. Res., 8, p.95-110.

Gronvold, K., 1972: Structural and petrochemical studies in the Kerlingarfjoll region, central Iceland. D.Phil. Thesis, Oxford University, 237 p.

Hambrey, M.J., and Harland, W.B., 1981: Earth's pre-Pleistocene Glacial Record. Cambridge University Press., 1004 p.

Hampton, M., 1972: Subaqueous debris in generating turbidity currents. J. Sedim. Patrol., 42, p.755-793.

Honnores, J., and Kirst, P., 1975: Submarine basaltic volcanism: Morphometric parameters for discriminating hyaloclastites from hyalotuffs. Bull. Volc., 39, p.1-25.

Imsland, P., 1978: The geology of the volcanic island Jan Mayen, Artic Ocean. Nord. Volcanol. Inst., Rep. 7813.

Jakobsson, S.P., 1979: Petrology of Recent basalts of the eastern volcanic zone, Iceland. Acta Nat. Islandica 26, 103 p.

Johannesson, H., Jakobsson, S.P., and Saemundsson, K., 1982: 1:250000 Geological map of Iceland, sheet 6, S-Iceland. Icelandic Museum of Natural History and Iceland Geodetic Survey, Reykjavik.

Johnson, A.M., 1970: Physical processes in geology. Freeman, Cooper and Co., San Francisco.

Jones, J.G., 1969: Pillow lavas as depth indicators. Am.

Journ. Sci., 267, sp.181-195.

Jones, J.G., 1970: Interglacial volcanoes of the Laugarvatn region, southwest Iceland, II: Journ. Geology, 78, p.127-140.

Jones, J.G., and Nelson, P.H.H., 1970: The flow of basalt from air into water - its structural expression and stratigraphic significance. Geol. Mag., 107, p.90-96.

Kjartansson, G., 1943: Arnesingasaga I: Natturulysing Arnessyslu, Yfirlit og jardsaga, Reykjavik, 268 p.

Kjartansson, G., 1959: The Moberg Formation; II, In: Thorarinsson, S., (ed.), On the geology and geomorphology of Iceland. Geografiska Annaler, 41, p.139-143.

Kjartansson, G., 1966: Sur la recession glaciare et les types volcaniques dans la region du Kjolur sur le plateau de l'islande. Rev. Geomorphol. Dyn., 16, p.23-39.

Larsen, G., 1984: Recent volcanic history of the Veidivötn fissure swarm, Southern Iceland - An approach to volcanic risk assessment. Journ. Volcan. Geoth. Res., 22, p.33-58.

Lowe, D.R., 1976: Subaqueous liquified and fluidized sediment flows and their deposits. Sedimentology, 23, p.285-308.

Lowe, D.R., 1982: Sediment gravity flows: II. Depositional models with special reference to the deposits of high-density turbidity currents. Journ. Sed. Petrol, 52, p.279-297.

Mac Donald, G.A., 1972: Volcanoes. Prentice Hall, Inc., Englewood Cliffs, N.J., 510 p.

McGowen, J.H., and Groat, C.G., 1971: Van Horn Sandstone, West Texas: An alluvial fan model for mineral exploration. Bureau Econ. Geol. Univ. of Texas at Austin, Report of investigations No. 72, 57 p.

Miall, A.D., 1978; Lithofacies types and vertical profile models in braided rivers; a summary. In: Fluvial Sedimentology, Miall,A.D., (ed.), Mem. Can. Soc. Petrol. Geol., 5, p.597-604.

Middleton, G.V., and Hampton, M.A. 1973: Sediment gravity flows: mechanism of flow and deposition. In: Turbidites and deep water sedimentation, Soc. Econ. Paleout. Minor., Pacific Sec., Short Course Lecture Notes, p.1-38.

Middleton, G.V., and Hampton, M.A., 1976: Subaqueous sediment transport and deposition by sediment gravity flows. In: Stanley, D.J., and Swift, D.J.P. (eds.)., Marine sediment transport and environmental management: New York, Wiley, p.197-218.

Moore, G.J., 1965: Petrology of deep-sea basalts near Hawaii. Am. J. Sci., 251, p.169-191.

Moore, J.G., Phillips, R.L., Grigg, R.W., Peterson, D.W., and Swanson, D.A., 1973: Flow of lava into the sea, 1969-1971 Kilauea Volcano Hawaii. Geol. Soc. Am. Bull., 84, p.537-546.

Nielsen, N., 1937: Vatnajokull: Kampen mellem ild og is. Copenhagen.

Niem, A.R., 1977: Mississippian pyroclastic flow and ash fall deposits in the deep-marine Ouachita flysch basin, Oklahoma and Arkansas. Geol. Soc. Amer. Bull., 88, p.49-61.

Noe-Nygaard, A., 1940: Subglacial volcanic activity in ancient and recent times (studies in the palagonite-system of Iceland No. 1). Folia Geographica Danica, Tom. I, no. 2, 67 p.

Peacock, M.A., 1962: The Volcano-glacial Palagonite Formation of Iceland. Geol. Mag., 63, p.385-399.

Pettijohn, F.J., Potter, P.E., and Siever, R., 1972: Sand and sandstone. Springer-Verlag N.Y., 618 p.

Pjetursson, H., 1900: The glacial palagonite-formation of Iceland. Scottish Geo. Mag., 16, p.265-293.

Preusser, H., 1976: The landscapes of Iceland, types and regions. Dr. W. Junkb.v., Publ., The Hague, Netherlands, 363 p.

Rittmann, A., 1960: Vulkane und Ihre Tatigkeit. Ferdinand Enke, Stuttgart, 336 p.

Saemundsson, K., 1967: Vulkanismus and Tektonik des Hengill- Gebietes in Sudwest-Island. Acta Nat. Islandica, 2, no. 7, p.1-105.

Saemundsson, K., 1970: Interglacial lava flows in the lowlands of Southern Iceland and the problem of two-tired columnar jointing. Jokull, 20, p.63-77.

Saemundsson, K., 1972: Notes on the geology of the Torfajökull central volcano. Natturufraedingurinn, 42, p.81-99.

Saemundsson, K., 1979: Outline of the geology of Iceland. Jokull, 29, p.7-28.

Sanders, J.E., 1965: Primary sedimentary structures formed by turbidity currents and related resedimentation mechanisms. In: Middleton, G.V. (ed.): Primary sedimentary structures and their hydrodynamic interpretation. Soc. Econ. Paleontol. and Mineral. Spec. Publ., 12, p.192-219.

Schmid, R., 1981: Descriptive nomenclature and classification of pyroclastic deposits and fragments: Recommendations of the IUGS subcommission on the systematics of igneous rocks. Geology, 9, p.41-43.

Schmincke, H.-U., 1982: Volcanic and chemical evolution of the Canary Islands. In: von Rad, U., Hinz, K., Sarnthein, M., and Seibold, E., (eds.), Geology of the Northwest African Continental Margin. Springer Verlag, Berlin, Heidelberg, New York, p.273-308.

Shaw, J., 1982: Melt-out till in the Edmonton area, Alberta, Canada. Can. Journ. Earth. Sci., 19, p.1548-1570.

Sheridan, M.F., 1979: Emplacement of pyroclastic flows: A review. In: Caphin, C.E., and Elston, W.E., (eds.), Ash flow tuffs. Geol. Soc. Am., Spec. Paper, 180, p.125-136.

Sigvaldason, G.E., 1968: Structure and products of subaquatic volcanoes in Iceland. Contr. Mineral. Petrol., 18, p.1-16.

Sparks, R.S.J., 1976: Grain size variations in ignimbrites and implications for the transport of pyroclastic flows. Sedimentology, 23, p.147-188.

Sparks, R.S.J., 1978: Gas release rates from pyroclastic flows: An assessment of the role of fluidization in their emplacement. Bull. Volcanol., 41, p.1-9.

Spry, A., 1962: The origin of columnar jointing, particularly in basalt flows. J. Geol. Soc., Australia, 8, p.191-217.

Swanson, D.A., 1972: Pahoehoe flows from the 1969-1971. Mouna Ulu eruption, Kilauea Volcano, Hawaii. Geol. Soc. Am. Bull., 84, p.615-626.

Thorarinsson, S., 1968: On the rate of lava- and tephra production and the upward migration of magma in four Icelandic eruptions. Geol. Rundschau, 57, p.705-718.

Thorarinsson, S., 1969: Glacier surges in Iceland, with special reference to the surges of Bruarjokull. Can. J. Earth Sci., 6, p.875-882.

Thoroddsen, T., 1925: Die Geschichte der Islandischen Vulkane. Det kgl. danske Vidensk. Selsk.skr. Nat. og Mat. Afdel. IX, 8, Kobenhavn.

Van Bemmelen, R.W., and Rutten, M.G., 1955: Table mountains of Northern Iceland. Leiden, E.J. Brill, 217 p.

Visser, J.N.J., 1983: Submarine debris flow deposits from the Upper Carboniferous Dwyka Tillite Formation in the Kalahari Basin, South Africa. Sedimentology, 30, p.511-523.

Walker, G.P.L., 1960: Zeolite zones and dike distribution in relation to the structure of the basalts of eastern Iceland. Journ. Geol. 68, p.515-528.

Walker, G.P.L., and Blake, D.H., 1966: The formation of a palagonite breccia mass beneath a valley glacier in Iceland. Quart. Jour. of the Geol. Soc. of London, 122, p.45-61.

Walker, R.G., 1975: Generalized facies models for resedimented conglomerates and turbidite associations. Geol. Soc. Am. Bull., 86, p.737-748.

Ward, G., 1974: The morphology of the Skeidara and Skaftafellsa river distributaries, Skeidararsandur, Iceland. Unpubl. B.A. thesis, Univ. of South Carolina, 91 p.

Waters, A.C., 1960: Determining direction of flow in basalts. Am. J. Sci., 258-A, p.350-366.

Wells, G., Bryan, W.B., and Pearce, T.H., 1979: Comparative morphology of ancient and modern pillow lavas. Journ. Geol., 87, p.427-440.

Williams, H., and McBirney, A., 1979: Volcanology. Freeman, Cooper and Co., San Francisco, 391 p.

Wilson, C.J.N., 1980: The role of fluidization in the emplacement of pyroclastic flows: An experimental approach. Journ. Volc. Geoth. Res., 8, p.231-249.

Wilson, C.J.N., and Walker, G.P.L., 1982: Ignimbrite depositional facies: The anatomy of a pyroclastic flow. J. Geol. Soc. London, 139, p.581-592.

Wohletz, K.H., and Sheridan, M.F., 1983: Hydrovolcanic explosions II. Evolution of basaltic tuff rings and tuff cones. Am. Journ. of Sci., 283, p.385-413.

Yokohama, S., 1974: Mode of movement and emplacement of Ito pyroclastic flow from Aira Caldera, Japan. Sci. Rep. Tokyo Kyoiku Daigaku, Sect. C, 12, p.17-62.