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The structure of the Cretaceous–Palaeogene sedimentary-volcanic area of Svartenhuk Halvø, central West Greenland

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Basalts, basin development, central West Greenland, Cretaceous–Palaeogene, extensional faults, Svartenhuk Halvø, transfer faults.

Cover
View of the 1160 m summit north-east of Simiuttap Kuua seen from the east; relief shown is about 1050 m. The view shows the striking contrast between the thick, brown-weathering flows of the Svartenhuk Formation forming the upper part of the mountain and the thinner, grey-weathering flows of the Vaigat Formation below. This contrast greatly facilitates mapping of the boundary between the formations over much of Svartenhuk Halvø. Near the bottom of the slope in the central part of the photograph, the contact between a cross-cutting dolerite sheet and the underlying sediments interbedded with hyaloclastite breccias can be seen. The poor outcrops at the foot of the slope to the right are of Precambrian gneiss.

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Frontispiece. **Above**: View of the western margin of the flexure zone between Usuit Kuussuat and Ulissat where it emerges at the coast 4 km south-west of Ulissat. Faulted lavas of the Vaigat Formation cross-cut by *c.* E–W-trending dykes. The height of the cliff on the right is *c.* 300 m. **Below**: View of part of the coast 3.5 km east of Tartuusaq showing lavas dipping south-west cut by dykes and faults dipping north-east. One fault (shown on Fig. 2) runs diagonally across the slope and has downthrow to the north-north-east. To the left (i.e. just left of centre of the picture) a dyke follows another fault that dips and throws down to the north-east. The height of the mountain on the right is 544 m.

Abstract

Larsen, J.G. & Pulvertaft, T.C.R. 2000: The structure of the Cretaceous–Palaeogene sedimentary-volcanic area of Svartenhuk Halvø, central West Greenland.
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Svartenhuk Halvø ('halvø' = peninsula) lies at the northern end of the exposed part of the Palaeogene volcanic province in central West Greenland. The peninsula is divided into two parts by a major NW–SE fault system, the Cretaceous boundary fault system. To the south-west lies a basinal area in which Cretaceous – Paleocene sediments are overlain by upper Paleocene basalts, whereas north-east of the fault system there is an elevated area in which Precambrian crystalline basement is exposed and is overlain by basalts, with only a few isolated pockets of Paleocene sediments in depressions in the basement surface.

Subsidence and sedimentation began in Albian time with the deposition of deltaic sandstones and mudstones. In the Turonian – early Campanian there was further subsidence and at least 1500 m of distal turbidites were deposited. During middle Campanian (?) – early Paleocene time the area was unstable, and alternating phases of uplift, erosion and subsidence resulted in the removal of all Cretaceous sediments in the uplifted Precambrian area, and the development of discordances in the basin area.

Volcanism started in mid-Paleocene time in a subsiding marine environment, so that the earliest volcanic rocks are hyaloclastite breccias. Later volcanism was almost entirely subaerial. The volcanic rocks are divided into two formations: (1) the Vaigat Formation (lower), dominated by picritic and other olivine-rich tholeiitic basalts, with a significant contaminated unit at the base of the formation in the south-east; (2) the Svartenhuk Formation (upper), characterised by plagioclase-porphyritic and aphyric basalts. In the northern part of the area about 50 m of fluvio-lacustrine sediments, tuffs and hyaloclastite separate the two formations. Mafic dykes occur throughout the area and are most numerous within the area of exposed Vaigat Formation. Thick sills and sheets of dolerite occur in the vicinity of the Cretaceous boundary fault system and within the sediments in the eastern part of the area.

Extensional faulting and tilting occurred in the basin area, both prior to, during and after volcanism. Most extensional faults trend NW–SE and throw down to the north-east, so that the rotated fault blocks dip south-west. However, there is a great difference in the degree of faulting and tilting between the northern and southern parts of the area, with few faults and generally low dips in the north, and numerous faults and dips between 20° and 40° in the south. The increase in the degree of extension and fault activity occurs at WNW–ESE- to E–W-trending transfer faults. Some extensional faults are associated with flexure zones with relative uplift on the north-east side and dips up to 60° to south-west within the zones. The regional extensional vector that gave rise to the extensional faulting could lie in any direction between NE–SW, normal to the extensional faults, and E–W, parallel to the transfer faults. The regional setting of the area suggests that extension was NE–SW. The regional structural pattern resembles closely (in mirror image) a structural pattern that has been described from the north-east side of the Gulf of Suez.

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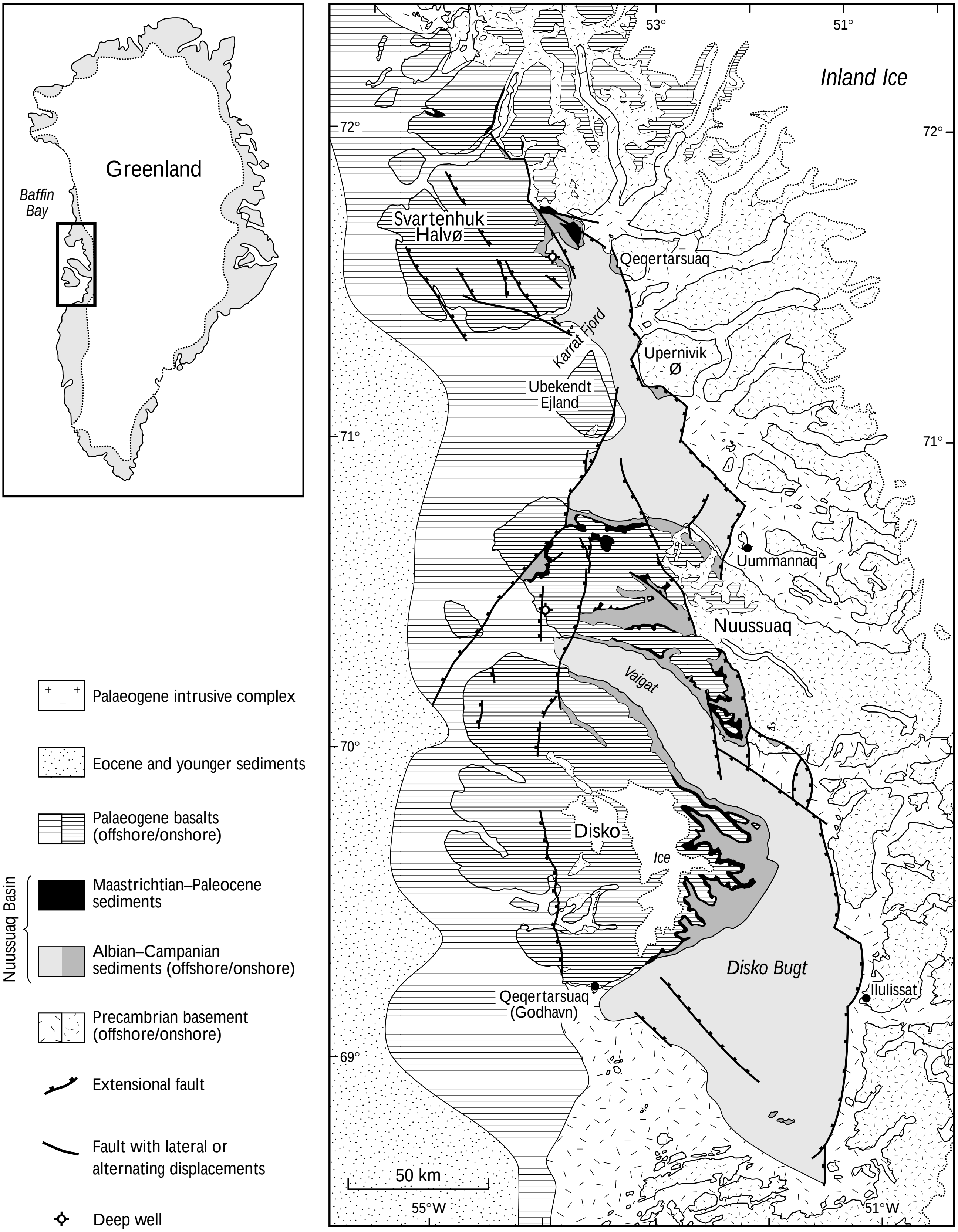

Fig. 1. Simplified map of the Nuussuaq Basin showing onshore and offshore geology and the position of Svartenhuk Halvø.

Introduction

Svartenhuk Halvø (‘halvø’ is Danish for peninsula) lies in the northern part of a substantial area of exposed Cretaceous–Palaeogene rocks in central West Greenland. This area extends from Disko Bugt (68°40′N) in the south through Nuussuaq to north of Svartenhuk Halvø (72°40′N) in the north (Fig. 1). It is the most extensive area of exposed Cretaceous–Palaeogene rocks in the entire Labrador Sea – Davis Strait – Baffin Bay region. Other areas of Cretaceous–Palaeogene rocks occur on the Bylot Island (Miall *et al.* 1980) and at Cape Dyer (Burden & Langille 1990), both on the Canadian side of Baffin Bay (see Fig. 14). In the Disko–Svartenhuk region the Cretaceous–Palaeogene is represented by fluvial, estuarine, deltaic, marine shoreface and marine shelf sediments of Albian to early Paleocene age that are overlain by late Paleocene – early Eocene continental flood basalts in both subaqueous and subaerial facies. In general the sediments pass from fluvio-deltaic sandstones, mudstones and coals in the south-east to distal marine turbidites in the north-west. The flood basalts overstep the margin of the sedimentary basin to lie directly on Precambrian basement to the east and north.

Previous investigations

The Disko–Svartenhuk region has been the subject of several phases of geological investigation since ammonites were first described from the area by Hoff (1865). Early work was concentrated on fossil plants (e.g. Heer 1883), invertebrate palaeontology (e.g. Ravn 1918) and coal (e.g. Steenstrup 1883). The first petrological studies of the volcanic rocks on Svartenhuk Halvø were carried out in the 1930s (Nieland 1931; Noe-Nygaard 1942). In the late 1940s and 1950s research was focused mainly on biostratigraphy and palaeontology (e.g. Birkelund 1965; Rosenkrantz 1970). In the late 1960s, however, interest in the Disko–Svartenhuk region took a new turn when the international oil industry recognised that the West Greenland shelf could have a potential for hydrocarbons. Sediments cropping out in nearby onshore areas therefore attracted attention as possible analogues of sedimentary environments and facies occurring in the shelf (Henderson 1969; Ehman *et al.* 1976). The results of studies carried out in the 1970s were summarised by Henderson *et al.* (1981), and mapping in the area resulted in the publication of five 1:100 000 geological map sheets, the northernmost two of which provide the basis of the present paper (Larsen 1983; Larsen & Grocott 1992).

Oil industry activity offshore culminated in the drilling of five exploration wells in 1976 and 1977. Results were however disappointing, and all licences were relinquished by early 1979. In the following decade studies in the Disko–Svartenhuk region were carried out largely by geologists from Copenhagen University. However, in 1990 it was decided that new petroleum-related investigations should be carried out in West Greenland, both onshore and offshore, with a view to reviving industry interest in the region, and the current cycle of multidisciplinary studies in the Cretaceous–Palaeogene outcrops onshore was initiated. When in 1992 oil and bitumen were discovered in vugs and vesicles in basalts in western Nuussuaq (Christiansen *et al.* 1994), these studies became more than analogue studies. The region was now a potential petroleum basin in its own right, and in 1994–1998 commercial exploration was carried out in Nuussuaq by the small Canadian company grønArctic Energy Inc. Six slim core wells have been drilled in Nuussuaq and Svartenhuk Halvø, four by grønArctic and two by the former Geological Survey of Greenland (GGU, amalgamated in 1995 with its Danish counterpart to become the Geological Survey of Denmark and Greenland – GEUS). Oil bled from the cores of two of these wells, traces of oil were found in cores of two other wells, and the cores of the remaining two wells, particularly that from the GGU well in eastern Svartenhuk Halvø (Umiivik-1), released wet gas (Bate & Christiansen 1996; Christiansen *et al.* 1996; Dam *et al.* 1998b). The search for oil in outcrops has resulted in seep discoveries over an area extending from northern Disko through western Nuussuaq to the south-east corner of Svartenhuk Halvø (Christiansen *et al.* 1998; Bojesen-Koefoed *et al.* 1999). In 1996 grønArctic drilled a conventional well to a depth of 2996 m in western Nuussuaq (Christiansen *et al.* 1997; Kristensen & Dam 1997).

In addition to oil discoveries, the current cycle of investigations in the Disko–Svartenhuk region has led to important advances in the understanding of the sedimentary history and basin development (Dam & Sønderholm 1994, 1998; Dam *et al.* 1998a). Until recently, however, little was known about the structure in the sedimentary basin, as only the structures affecting the flood basalts on Nuussuaq had been described in any detail (Pedersen *et al.* 1993; A.K. Pedersen, manuscript maps). The lack of understanding of the deep structure of the basin became apparent when GGU acquired a 13 km long seismic line on the south coast of Nuussuaq in 1994. This revealed that the thickness of sediments here is at least 6 km and perhaps 8 km, which is far greater than previously imagined (Christiansen *et al.* 1995). This discovery motivated a marine multichannel seismic programme that was carried out in 1995 in Disko Bugt, west of Disko, and in the fjords north and south of Nuussuaq. The results of this survey, together with gravity modelling and a compilation of onshore structures on Disko and Nuussuaq, are described in full by Chalmers *et al.* (1999). Since the marine seismic survey was not extended north of Ubekendt Ejland, Svartenhuk Halvø is not discussed in that paper.

Scope of this bulletin

The purpose of the present paper is to complement the papers by Chalmers *et al.* (1998, 1999), so that together the three papers provide an overall view of the structures in the Cretaceous–Palaeogene of central West Greenland as known today. The basis for the paper is the mapping by J.G.L. and his group carried out over four field seasons between 1974 and 1983; this work, which incorporated earlier reconnaissance mapping by Rosenkrantz *et al.* (1942), Münther (1973; mapping from 1952), Pulvertaft & Clarke (1966) and Rosenkrantz & Pulvertaft (1969), resulted in two Survey 1:100 000 map sheets: Igdlorssuit, 71 V.1 Syd (Larsen 1983) and Svartenhuk Halvø, 71 V.1 Nord (Larsen & Grocott 1992). These map sheets should be consulted for features such as dykes, sills and minor faults and fractures that could not be represented on Fig. 2 for reasons of scale.

Geological units

Svartenhuk Halvø is divided into two fundamentally different areas. The north-eastern part of the peninsula is an area of elevated Precambrian basement overlain by upper Paleocene flood basalts, with only local pockets of sediments which are entirely of late Paleocene age. In contrast, the south-western area is dominated by flood basalts which overlie sediments of Cretaceous and early Paleocene age; this area is referred to as the *basin area*. Whereas the basement area is structurally simple and largely unaffected by faults, all but the north-eastern part of the basin area is dissected by numerous faults and also zones with relatively steep dips. The two areas are separated by an irregular fault system which crosses the peninsula in a roughly NW–SE direction. The fault system is part of what is referred to as the *Cretaceous boundary fault system* in central West Greenland, because it defines the eastern and north-eastern boundary of present-day outcrops of Cretaceous sediments; it is believed that the area of Cretaceous sediments originally extended east of the fault system, and that Cretaceous sediments deposited east of the boundary fault line were removed by erosion during latest Cretaceous and early Paleocene time.

Nine geological units have been distinguished in Fig. 2 in order to illustrate the bedrock geology of the area and the structural pattern; in addition, some of the largest areas covered by Holocene fluvio-glacial deposits and ice have been shown, not least because fluvio-glacial deposits conceal some of the major faults, leaving their position as a matter of conjecture. Several more units are distinguished on the 1:100 000 map sheets, and a full description of the stratigraphy of the volcanic rocks is in preparation. A summary of the rock units and tectonic events described in this bulletin is presented in Fig. 3.

Precambrian basement

Precambrian basement outcrops in the north-east part of the area. The main units in the Precambrian are

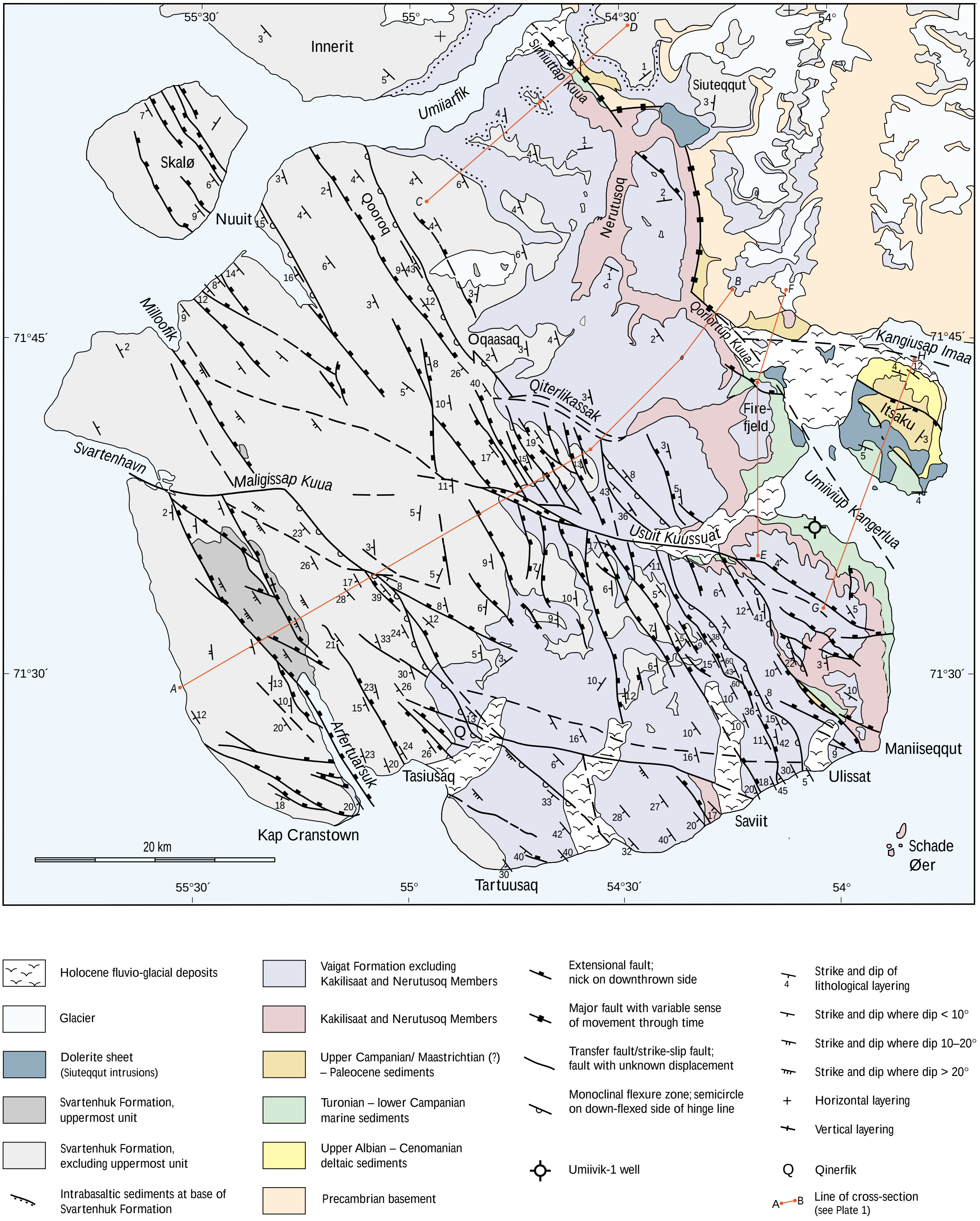

Fig. 2. Simplified geological map of Svartenhuk Halvø. For location, see Fig. 1

Archaean gneisses and Palaeoproterozoic supracrustal rocks, mainly schists and metamorphosed turbidite sandstones and mudstones, which together form part of the Rinkian (1.9–1.7 Ma) structural province in northern and central West Greenland (Henderson & Pulvertaft 1987). The structural grain (trend of foliation and fold axes) in the basement is N–S in the area north-west of the inner end of Kangiusap Imaa, swinging to NW–SE in the northern part of the area (van den Eeckhout & Grocott 1982; Grocott & Vissers 1984). Later fractures trend both NNW–SSE and WNW–ESE.

The Precambrian basement is overlain by flood basalts belonging to the uppermost part of the lower basalt formation (Vaigat Formation) and the more extensive upper basalt formation (Svartenhuk Formation). The pre-basalt basement surface is an undulating surface rising from 200 m above sea level (a.s.l.) on the north-west side of Umiiarfik to 900 m in the east, with local palaeomountains rising 700 m above this surface. Locally, in topographic lows in the basement surface, fluvial sediments with thin coals lie between the basement and the overlying basalts.

Lower–mid-Cretaceous deltaic and marginal marine sediments

These are only exposed on Itsaku peninsula in the eastern part of the area. The succession begins with a basal conglomerate in which the clasts are entirely of metamorphic rocks; this conglomerate lies unconformably on weathered Precambrian metasediments. The conglomerate is succeeded by *c.* 350 m of sediments arranged in coarsening-upwards cycles of mudstone, sandstone (the dominant lithology) and thin coal seams. These cycles are similar to those in the Atane Formation on Nuussuaq (Pedersen & Pulvertaft 1992) and were deposited from prograding wave- and storm-influenced deltas (F. Dalhoff and G. Dam, personal communication 1999). The cross-bedded sandstones that top the deltaic cycles were deposited in straight to slightly sinuous channels; foreset dip directions vary from west through north-west to just east of north, with maximum towards 305°. The deltaic succession is overlain by *c.* 70 m dark mudstones with abundant plant remains. Because of the thermal influence of the many dolerite sills on Itsaku, palynomorphs have been destroyed throughout most of the section. However, in the lowest 70 m of the section there is a pollen assemblage that bears affinity to the Late Albian – Early Cenomanian assemblages from the Cretaceous sediments on the islands Qeqertarsuaq and Upernivik Ø to the east and south-east (Croxton 1978a). The overlying dark mudstones have yielded blackened pollen assemblages that could well be of Late Cretaceous age (Croxton 1978a).

This part of the sedimentary section on Itsaku terminates at the base of a conglomerate horizon which is discussed later (see p. 12).

Turonian – lower Campanian marine mudstones and distal turbidites

Dark mudstones with thin turbiditic sandstone layers, laminae and stringers are exposed in many stream gullies in the eastern part of Svartenhuk Halvø and also in low coastal outcrops around the bay Umiiviup Kangerlua. These sediments have been studied in the core of slim core well Umiivik-1 drilled on the south coast of this bay (Dam 1997; Dam *et al.* 1998b), and they show little variation throughout the 1200 m cored (240 m of which consists however of dolerite sills). In this core there are three main cycles, each with a weak thickening- and coarsening-upwards tendency; the uppermost cycle is overlain by a unit lacking a coarsening-upwards pattern but containing a relatively large amount of slump deposits. The environment of deposition was a distal marine slope (Dam 1997; Dam *et al.* 1998b). Similar mudstones and thin turbiditic sandstones have also been cored in a number of shallow wells (maximum depth 86 m) in eastern Svartenhuk Halvø. It is significant that even in the core drilled only 1.6 km from a 700 m high, steep slope of Precambrian gneisses, the sediments retain their distal marine character. This implies that when the Turonian – lower Campanian sediments were deposited, this drill site did not lie close to a steep coast and that these sediments originally continued across the site of the boundary fault system into the area to the north-east.

The age of the Turonian – lower Campanian sediments is fairly accurately known from ammonites and dinoflagellates (Birkelund 1956; Nøhr-Hansen 1996, 1997). The oldest dinoflagellates occur at 540 m below surface in the Umiivik-1 well and indicate a Turonian age (dinoflagellates in the mudstones below this have been destroyed by heating from sills), while the youngest dinoflagellate zone, which occurs immediately below hyaloclastite basalt breccias on the west side of Umiiviup Kangerlua, is late Santonian/?early Campanian. Younger marine mudstones occur on the south side of Itsaku, where ammonites indicating an early

Period		Sedimentary and volcanic formations	Tectonic events
		Fluvio-glacial sediments	
Pleistocene		Moraines	Post-glacial isostatic uplift
		Ice caps and glaciers	
---1.8 Ma---			
Neogene			↕ ? Regional uplift
---24 Ma---			
			↕ ? Inversion south of Simiuttap Kuua
	↕		Extensional faulting; reactivation of basement faults
Palaeogene	dykes — sheets	Uppermost Svartenhuk Formation basalts	
		Arfertuarsuk trachyte	
		Svartenhuk Formation: plag.-porph. basalts	
		Fluvio-lacustrine sediments (Simiuttap Kuua area)	
		Vaigat Formation: picritic basalts – subaerial	
61 Ma		Vaigat Formation: " " – subaqueous;	↕ Extensional faulting
		Kakilisaat and Nerutusoq Members at base	Subsidence
		Conglomerate and mudstone (Itsaku, Firefjeld and NE of Qorlortup Kuua)	
--- 65 Ma ---			Differential uplift and erosion
		Conglomerate and mudstone (Itsaku)	Subsidence
			Uplift and erosion
Late Cretaceous		Distal turbidites (Umiiviup Kangerlua and Firefjeld; S side Simiuttap Kuua)	Thermal subsidence
		Mudstones (Itsaku)	
--- 99 Ma ---		Deltaic sediments (Itsaku)	↕ Rifting
Early Cretaceous			

Fig. 3. Summary of late Phanerozoic rock units and tectonic events on Svartenhuk Halvø. The vertical scale is not proportional to time.

Campanian age have been collected (Birkelund 1965). Since hyaloclastite breccias in some places lie directly on lower Santonian marine mudstones, the sediment-hyaloclastite contact reflects a substantial hiatus and is probably a discordant erosion surface.

Marine mudstones interbedded with thin sandstones outcrop in stream gullies on the south side of Simiuttap Kuua in north-west Svartenhuk Halvø ('kuua' is Greenlandic for river). A conglomerate layer consisting of slabs of sandstone up to 80 cm long and 30 cm thick lying with no preferred orientation in mudstones indicates that there was intermittent tectonic activity during the deposition of these sediments (Ehman *et al.* 1976). Unfortunately the age of these sediments cannot be determined closer than Cenomanian – early Campanian, which is the age range of belemnites collected from this locality (Birkelund 1956).

Mudstones and siltstones dated as Coniacian–Santonian (H. Nøhr-Hansen, personal communication 1998) occur 4 km north-west of Maniiseqqut. This occurrence is bounded by NW–SE faults, and the sediments are folded on axes parallel to the faults; axial surfaces of the folds are both steep and flat-lying. Folds with both steep and low-dipping axial surfaces are also seen in outcrops of Cretaceous sediments on the south side of the Usuit Kuussuat valley. This folding does not affect the overlying basalts, but it is not clear whether the folds were generated by gravity gliding triggered by fault movements or were the response of these ductile rocks to post-basalt fault movements.

Middle Campanian/Maastrichtian (?) – lower Paleocene mudstones and conglomerates

The summit of the Itsaku peninsula consists of dark mudstones which have yielded pollen indicating a Paleocene age (Croxton 1978a). Lower down, on the east side of the peninsula, there are two conspicuous conglomerate horizons, one at about 300 m above sea level and the other 150–200 m higher up (Plate 1). The lower conglomerate is 20 m thick in the south, thickening northwards to 70 m. A low-angle unconformity separates this from the underlying sediments. The conglomerate consists of fining-upwards beds in which the lower parts contain rounded clasts of metamorphic rocks, quartz and weathered basic igneous rocks up to 1 m in size. The upper conglomerate is poorly exposed, but can in places be studied on the north-east slope at about 670 m a.s.l., where it contains rounded boulders and cobbles of sandstone, basement lithologies, and dark fine-grained rocks that have been not yet been properly described (Ehman *et al.* 1976; Croxton 1978a). No palynomorphs or other fossils have been recorded from the mudstones between the conglomerates, so the age of these sediments can only be inferred by analogy with Nuussuaq. Here there are similar conglomerates of both middle Campanian and middle Maastrichtian age, both of which overlie unconformities and fill incised valleys and submarine canyons (Dam & Sønderholm 1994; Dam *et al.* 2000). Thus the lower conglomerate on Itsaku could be middle Campanian and the upper one Maastrichtian, or alternatively the two levels of conglomerate could be the equivalent of the Maastrichtian and early Paleocene channel-fill conglomerates and sandstones in south-western Nuussuaq (Dam & Sønderholm 1998). Dam *et al.* (1998a) suggest that these Paleocene features on Nuussuaq reflect phases of uplift, erosion and valley incision related to the arrival of a mantle plume, a phenomenon that could be expected to have affected an extensive area of central West Greenland including Svartenhuk Halvø.

On the north side of Qorlortup Kuua there is an outcrop of dark mudstones and siltstones which lie on coarse conglomerates, below which there is an unknown thickness of mudstones. The boulders in the conglomerate are angular, up to 5 m in size, and consist entirely of basement lithologies. The conglomerates are arranged in fining-upwards cycles and grade into sandstones; large boulders can however occur sporadically in the upper parts of the conglomerate beds (G. Dam, personal communication 1999). In the mudstones immediately overlying the conglomerates Croxton (1978b) recovered specimens of *Alnipollenites*, which is indicative of a Paleocene age. The mudstones and conglomerates onlap a gneiss slope which dips south at up to about 30°; in places mudstones fill palaeovalleys eroded into the steep gneiss slope. This steep gneiss slope is regarded as an eroded and bevelled fault plane (see p. 20). Hyaloclastite breccia fills a palaeovalley in the mudstones.

On the south-east corner of Firefjeld a 30 m thick pebble and cobble conglomerate overlain by sand occurs immediately below the hyaloclastite breccia. On aerial photographs this bed can be traced northwards, and 4.5 km to the north the horizon is represented by a 5 m thick fining-upwards sandy bed with cobbles. The cobbles consist mainly of basement lithologies but also lithified conglomerate clasts occur. This conglomerate may be a correlative of the conglomerate north of Qorlortup Kuua.

About 5.5 km north-west of Maniiseqqut there is a small outcrop of Paleocene sediments dipping 15° to south-south-west. The sediments consist of sandstone grading up into black tuffaceous sandstone with molluscs and *Ophiomorpha* burrows, which in turn is overlain by hyaloclastite breccia. On the basis of dinoflagellate cysts Hansen (1980) placed these sediments in the upper part of the lower Danian. The contact between these rocks and the basalts to the south-west is a fault.

Volcanic rocks

The volcanic rocks on Svartenhuk Halvø are divided into two formations: the *Vaigat Formation* (lower) and the *Svartenhuk Formation*; each formation is divided into a number of members, of which only three have been distinguished on Fig. 2, two with the same colour. A full description of the stratigraphy of the basalts by J.G.L. is in preparation.

Throughout Svartenhuk Halvø the Vaigat Formation is overlain by the Svartenhuk Formation (Plate 1). At a few localities, e.g. north-east of Simiuttap Kuua and south of Usuit Kuussuat, there is evidence of angular unconformity between the Svartenhuk and Vaigat Formations, but over much of the peninsula there is no obvious unconformity between the Svartenhuk Formation and the Vaigat Formation.

Vaigat Formation

The type locality of the Vaigat Formation is the sound Vaigat which separates Disko from Nuussuaq peninsula (Hald & Pedersen 1975). It is the lowermost volcanic formation in the type area, just as it is on Svartenhuk Halvø. The Vaigat Formation in the type area has yielded ^{40}Ar/^{39}Ar ages between 60.7 ± 0.5 and 60.3 ± 1.0 Ma (Storey *et al.* 1998), but the oldest lavas here show normal magnetisation (Riisager & Abrahamsen 1999) which means that they must have been erupted during magnetochron 27n or between 61.3 and 60.9 Ma in the time scale of Cande & Kent (1995). This is the time when sea-floor spreading started in the inner Labrador Sea (Chalmers & Laursen 1995) and most likely also in Baffin Bay (Chalmers & Pulvertaft in press).

The Vaigat Formation is characterised by tholeiitic picrites and other olivine-rich basalts (Clarke & Pedersen 1976). On both Disko and Nuussuaq, horizons of orange-brown- or greyish brown-weathering silica-enriched, contaminated basalt are interspersed within the olivine-rich basalts. On Svartenhuk Halvø similar brown-weathering, silica-enriched, contaminated rocks occur at the base of the formation in the south-eastern part of the area and these are distinguished as the *Kakilisaat Member*. In the north-west a different brown-weathering unit occurs at the base of the Vaigat Formation; this consists of uncontaminated olivine basalts and is called the *Nerutusoq Member*. On Disko and Nuussuaq the earliest lavas in the Vaigat Formation were erupted at the sea floor and consist of pillow lava and hyaloclastite mounds. In due course the volcanic edifice emerged above sea level but to the east and north a subaqueous environment persisted, first as a marine embayment, later as a lake. On reaching the shoreline the subaerially erupted lavas passed into subaqueous hyaloclastite breccias which built up very large prograding Gilbert-type deltas with foresets reaching a maximum of 700 m in height on Nuussuaq (Pedersen *et al.* 1993). The heights of the foresets are a measure of the depth of the basin into which Vaigat Formation lavas flowed, or in other words the minimum pulses of subsidence prior to and during initial volcanism.

On Svartenhuk Halvø the earliest eruptions in the eastern and northern parts of the basin area also took place in a submarine environment, giving rise to hyaloclastite breccias and pillow lavas. As on Nuussuaq, the hyaloclastites form Gilbert-type delta structures, with foresets up to 250 m high in the upper, picritic breccias. Water depths during eruption of the lower breccias may have been greater, but these breccias are less well exposed and their foreset bedding is less distinct. In the lower part of the breccias there is some interfingering between breccia and mudstone.

Foreset dip directions in hyaloclastite breccias are an indication of flow direction and, in favourable situations, can be used to locate centres of eruption. Foreset directions recorded on Svartenhuk Halvø indicate that there was a centre of eruption at the north-west end of Umiiviup Kangerlua from which hyaloclastites spread towards the west, and another centre close to the Cretaceous boundary fault north-east of Qiterlikassak from which hyaloclastites spread in almost all directions but particularly towards the south (Larsen 1981a). Yet another eruptive site was situated in the Nerutusoq valley, from which the Nerutusoq Member was erupted and partially blocked the north-westerly spread of the picrites. Otherwise the general flow direction during subaqueous eruption was towards north and north-east. Feeder dykes have been observed in the hyaloclastite breccias both north-west of Firefjeld and north of Kangiusap Imaa (Larsen 1981a).

The thickness of the hyaloclastite breccias and pillow lavas on Svartenhuk Halvø increases from south to north and reaches a maximum of at least 600 m along the north-east side of the basin area. Between Maniiseqqut and Umiiviup Kangerlua cross-bedded hyaloclastites alternate with subaerial lavas, indicating that subsidence here was gradual and more or less kept pace with growth of the volcanic pile. There is also evidence in the area between Usuit Kuussuat and Maniiseqqut that north-east-facing fault scarps existed during eruption of the lower part of the Vaigat Formation. Here one can see examples of subaerial lava flows that pass north-eastwards into hyaloclastites as they cross WNW–ESE or NW–SE faults, the base of the flow dropping abruptly as it crosses the fault and the rocks become hyaloclastites.

When the subsided basin became filled by hyaloclastite breccias and subaerial lavas, the lavas overstepped the Cretaceous boundary fault system and flowed northwards and north-eastwards over the Precambrian basement surface. At times, however, either phases of renewed subsidence led to marine transgressions or, in the later stages of volcanism, the lavas dammed up rivers so that ephemeral lakes developed. In either case, on entering the subaqueous environment the lavas formed hyaloclastite breccias now seen as horizons within the predominantly subaerial lavas that overlie the basement in north-eastern Svartenhuk Halvø. To the south-west the exposed part of the Vaigat Formation is entirely subaerial. The interaction between volcanism and lake development has been documented in detail on Nuussuaq and Disko (G.K. Pedersen 1989; A.K. Pedersen *et al.* 1996; G.K. Pedersen *et al.* 1998a, b).

The Kakilisaat Member is an important unit in the structural interpretation of southern Svartenhuk Halvø. As already mentioned, it occurs at or near the base of the Vaigat Formation and consists of silica-enriched, contaminated basalts including olivine-bearing types; these are mainly developed in subaqueous facies in the eastern part of the basin (the 'brown breccia' of S. Munch in: Rosenkrantz *et al.* 1942), but in southern Svartenhuk Halvø, like the rest of the Vaigat Formation here, they occur in subaerial facies. Due to faulting, the outcrops of these contaminated basalts on Svartenhuk Halvø are not continuous, but the geochemical and petrological similarity of these rocks, regardless of where they occur or whether they are in subaqueous or subaerial facies (J.G. Larsen, unpublished data), suggests strongly that they belong to a single contiguous unit. Furthermore, over a large area of south-east Svartenhuk Halvø the contaminated lavas are overlain by a characteristic massive picrite flow with large olivine phenocrysts, which suggests that all the contaminated lavas in this area belong to the same stratigraphic unit.

Chemical analyses of the Kakilisaat Member show close similarities with analyses of the Kûgánguaq and Tunoqqu Members which occur within the Vaigat Formation on Disko and Nuussuaq respectively (Pedersen 1985a, b; Pedersen *et al.* 1996). However, correlation at this level of resolution must await the results of the palaeomagnetic investigations that were initiated on Svartenhuk Halvø in 1999 (Christiansen *et al.* 2000).

The thickness of the Vaigat Formation as a whole (including the Kakilisaat and Nerutusoq Members) increases from north to south. South-west of Simiuttap Kuua the total thickness of the unit is 950–1000 m, of which 600 m consist of hyaloclastite breccias. West of Qorlortup Kuua the hyaloclastites are about 500 m thick, and the thickness of the overlying subaerial lavas has increased to 600 m. North and south of Usuit Kuussuat thicknesses are harder to estimate because of displacements on faults and fractures zones that cannot be satisfactorily quantified. However, it appears that here the subaerial component of the Vaigat Formation is more than a kilometre thick, whereas south-west of Umiiviup Kangerlua the hyaloclastite breccias are only 300–350 m thick.

It is difficult to estimate the total thickness of the Vaigat Formation along the south coast of Svartenhuk Halvø. Subaerial lavas of the formation outcrop along the entire coast from Ulissat to Tartuusaq. Along the stretch from Saviit to Tartuusaq the width of outcrop in the direction of dip is about 17.5 km and the dip is consistently to the south-west at angles between 17° and 40°. This would imply that the Vaigat Formation here is about 9 km thick as suggested by Noe-Nygaard (1942), i.e. that there is an approximately five-fold increase in the thickness of the Vaigat Formation between latitude 71°45′N and the south coast of the peninsula. This is unlikely, as already pointed out by Münther (1973) and Larsen (1981b), the latter suggesting that the thickness is of the order of 4–4.5 km. There are at least two possible explanations.

The first is that there was continuous flexure and subsidence in southern Svartenhuk Halvø during the eruption of the lavas, and that, as the lavas poured out, both the volcanic sources and the area of subsidence migrated westwards. Such a process can result in stratal shingling analogous to that seen in some pull-apart basins where stratigraphic thicknesses greatly exceed the true thickness at any point in the basin (Crowell 1982, figs 6, 7). The alternative explanation, which is based on the established fault pattern in the area, is that the south-westward dipping Vaigat Formation in southern Svartenhuk Halvø is not an uninterrupted succession but has been repeated several times along faults with downthrow to the north-east. The two explanations are not mutually exclusive, and both syn- and post-volcanic faulting can have affected a shingled succession. The problem will be discussed in the final section of this paper (see p. 34).

Svartenhuk Formation

The Svartenhuk Formation consists mainly of brownish or greyish weathering tholeiites with or without olivine; these tholeiites are commonly plagioclase-porphyritic but can also be aphyric. One significant unit in the formation is dark grey-green and olivine-porphyritic. These rocks are the stratigraphic equivalent of the Maligât Formation on Disko and Nuussuaq (Hald & Pedersen 1975). The Svartenhuk Formation is at least 2800 m thick. Flows in this formation tend to be thicker than in the Vaigat Formation, with entablature giving rise to ‘trap’ morphology. Near the top of the Svartenhuk Formation there is a conspicuous anorthoclase-porphyritic trachyte, the Arfertuarsuk trachyte (Nieland 1931; Plate 1), which is also noteworthy in that it contains numerous baked mudstone inclusions (A.K. Pedersen, personal communication 1999). In order to bring out the structure of the area, the Arfertuarsuk trachyte and the basalts above are distinguished from the remainder of the Svartenhuk Formation in the map Fig. 2.

In the area north-east and south-west of Simiuttap Kuua and north-west of Umiiarfik the lowermost basalts of the Svartenhuk Formation are separated from the underlying olivine-rich basalts of the Vaigat Formation by about 50 m of hyaloclastite breccias, water-lain tuffs, and fluvio-lacustrine sandstones interbedded with dark mudstones and coal seams (Fig. 2; Plate 1). These mark an interval during which this part of the area lay at the fringe of volcanic influence. Similar sediments occur farther to the north-east where they occupy depressions in the basement surface. The petrography of the sandstones shows that most of the material was brought down from exposed basement terrain. The first lavas of the Svartenhuk Formation that reached this area flowed into lakes and hence are in subaqueous facies. After this subaerial conditions prevailed, and flood basalts spread over a very large area, reaching beyond the margin of the Inland Ice to the east.

Minor intrusions

Dykes

Mafic dykes occur throughout the area. An impression of the distribution and directions of the dykes can be gained from Fig. 4, which shows as many of the dykes recorded on the 1:100 000 sheets as the scale of the figure permits. However, as explained later, most of the dykes shown in the area of exposed basalts were recorded from aerial photographs, and there is a bias towards dykes which show the greatest resistance to erosion relative to their surroundings.

In the field, dykes can be divided into four main groups: picritic, olivine-porphyritic, plagioclase-porphyritic and aphyric basalts. These correspond to the four main lava flow types that can be identified in the field, and it is assumed that dykes and lava flows of similar composition belong to the same phase of volcanism, and that many dykes are feeders to the corresponding flows.

Dykes present in the lowermost part of the Vaigat Formation that have the same appearance and chemical composition as the contaminated basalts of the Kakilisaat Member and the overlying olivine basalts are believed to be the oldest exposed dykes, because the characteristic compositions of these units are not repeated in the younger lava units. These dykes are poorly exposed, and we have no systematic data concerning their orientation.

Picritic dykes corresponding to the upper part of the Vaigat Formation are commonest in the southern part of the area. They are generally rather thin, *c.* 0.4–1 m, although master dykes several metres thick occur, some of which are up to 25 m thick. These dykes are eroded just as severely as the host picrite lavas, and the colours of weathered surfaces are the same, so that these dykes are difficult to trace on aerial photographs and only very few are included in Figs 4 and 5. They have trends mainly between NE–SW and E–W, but also NW–SE and NNW–SSE trends have been observed. In several cases picritic dykes can be seen to change direction.

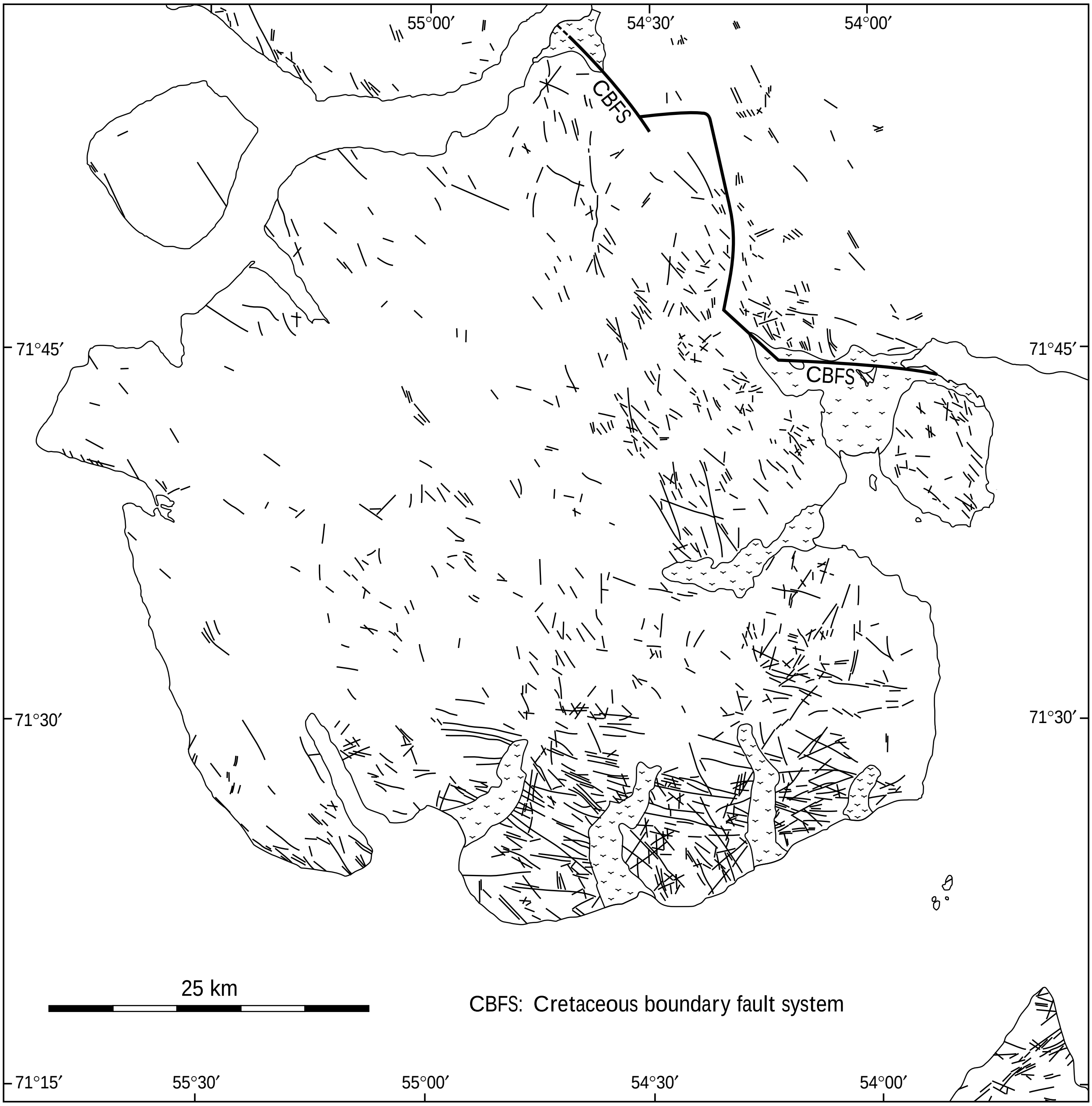

Fig. 4. Map showing dykes recorded on the 1:100 000 map sheets Svartenhuk, 71 V.1 Nord and Igdlorssuit, 71 V.1 Syd. Main areas of Holocene deposits are shown. For location, see Fig. 1.

Some NW-trending picritic dykes exposed in the steeply dipping lavas in southern Svartenhuk Halvø dip northeastwards, suggesting that they were tilted together with the lava pile. Feeder dykes with E–W and NW–SE trends have been observed in the central eastern area.

Olivine-poor, plagioclase-porphyritic and aphyric dykes cutting the Vaigat Formation are believed to be contemporaneous with equivalent lavas in the Svartenhuk Formation, although some may be feeders to younger lavas that have been removed by erosion. These dykes stand out as resistant walls in the crumbling picritic and olivine-rich flows of the Vaigat Formation and are therefore easy to trace on aerial photographs, but within the Svartenhuk Formation there is either no topographic expression of these dykes or they may form negative features and be indistinguishable from fault and fracture features.

A photogrammetric study of the orientation of the

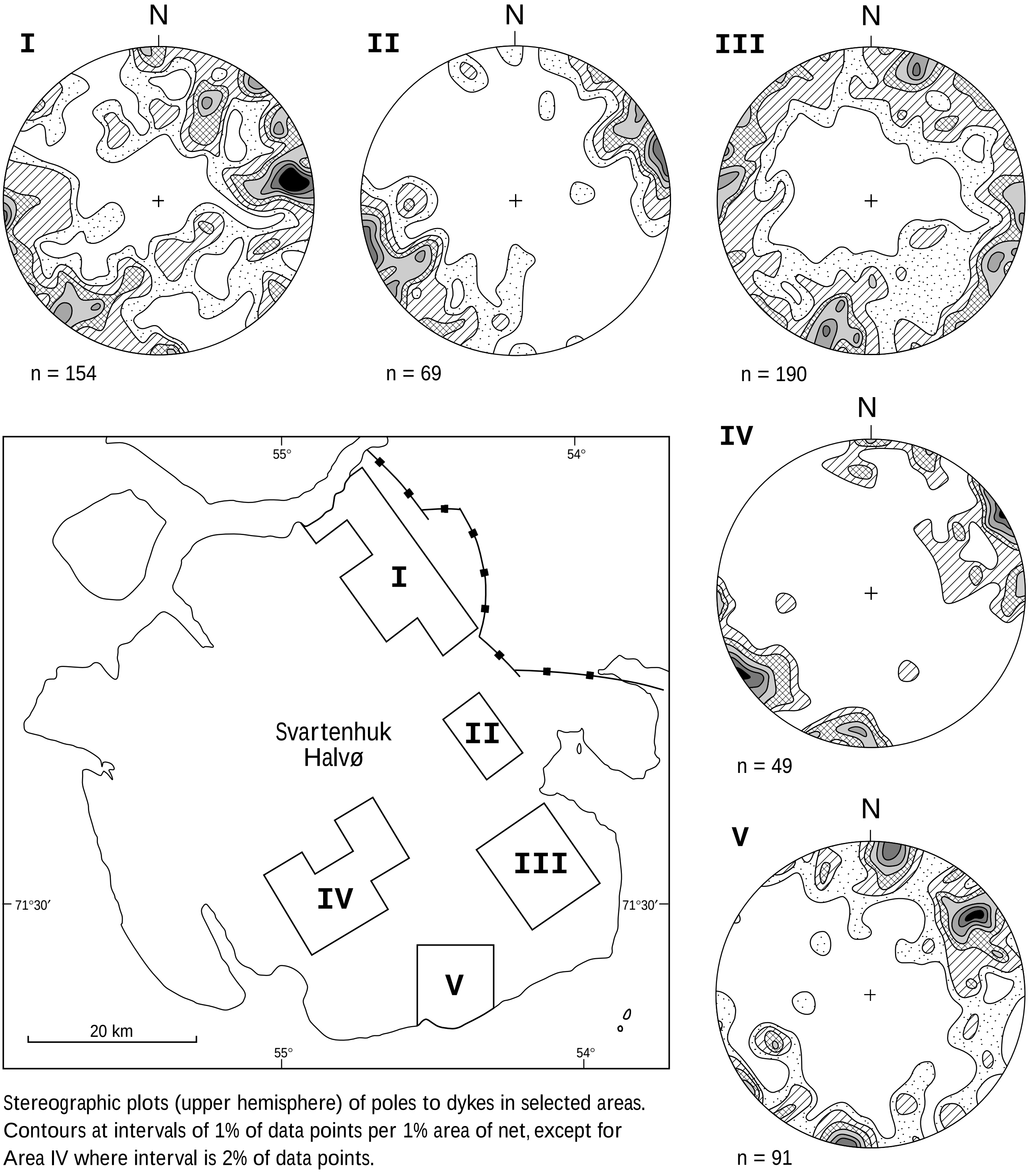

Fig. 5. Stereographic plots of poles to dykes in selected areas. The poles were plotted in the upper hemisphere so that maxima would lie in the same quadrant as the direction of dip.

dykes has been carried out in five areas which were selected to give an impression of regional variation in trend and dip patterns and how these are related to fault patterns and dip of host lavas. For practical reasons four of the five areas lie entirely within the area of outcrop of the Vaigat Formation where dykes are most easily seen in aerial photographs. In the Svartenhuk Formation only a small minority of dykes can be

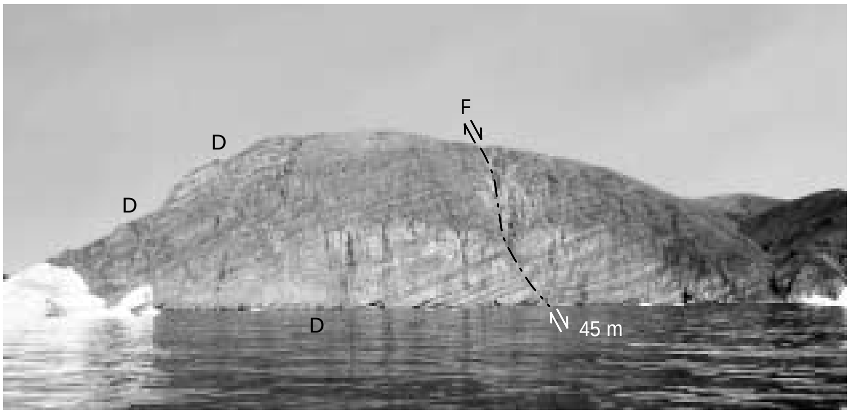

Fig. 6. View of a cliff west of Tasiusaq, showing basalts of the Svartenhuk Formation dipping south-west, cut by two dykes (**D**) and an extensional fault dipping north-east. Whether the dykes were tilted together with the basalt lavas or intruded into already dipping fractures has not been ascertained here. The height of the cliff is 300 m. For location, see Fig. 2.

seen on aerial photographs, and only in one of the selected areas, Area IV (Fig. 5), is there a substantial outcrop of Svartenhuk Formation. This method is adequate for providing a general indication of dyke orientations in an area, but is not completely reliable for determining dyke dips; a real change in dyke trend on a ridge or in a valley can be erroneously interpreted as interference between the surface of the terrain and a dipping dyke.

As can be seen from Figs 4 and 5, different areas are dominated by different trends. In the south, WNW–ESE to WSW–ENE trends dominate, together with NE–SW trends in the south-eastern area and some NW–SE- and N–S-trending dykes in the south-west. In the rest of the peninsula NW–SE and N–S trends dominate, while WNW–ESE-trending dykes form a minor group. As already pointed out, the vast majority of the dykes recorded are olivine-poor dykes post-dating the Vaigat Formation. The degree of scatter in both trends and dips of recorded dykes can best be seen from the contoured stereographic plots of poles to dykes in selected areas (Fig. 5).

In Area I, the northernmost area, dips in the basalts are less than 6° to SW. In spite of considerable scatter, there are distinct dyke maxima in trend 174°, dip 77° E and in 130/approximately vertical. These trends are approximately parallel to the trends of segments of the Cretaceous boundary fault system in this area. In Area II, where basalt dips are also very low, there is a clear dominance of *c.* NW–SE-trending dykes, approximately parallel to the Qorlortup Kuua segment of the boundary fault system and to the flexure zones (see p. 20).

In Area III in the south-east the lavas dip up to 12° SW, except locally in flexure zones where dips can exceed 45° SW (see p. 27). The scatter shown by the stereographic plot is considerable. The maximum trend is in 108°, which is approximately parallel to the transfer faults in southern Svartenhuk (see p. 29). There is also a concentration of trends in NNE–SSW.

Unlike Areas I–III, Area IV includes a large area of outcrop of the Svartenhuk Formation. As explained already, dykes are less easy to see in aerial photographs in the Svartenhuk Formation than in the Vaigat Formation, which is one reason for the low number of observations from this area. The stereographic plot shows a maximum concentration of dyke trends in 148°, which is also the dominant direction of extensional faults on Svartenhuk Halvø. There are also several dykes trending between E–W and ESE–WNW, parallel to the direction of transfer faults.

In Area V the dip of the lavas is generally steeper than 25° SW and locally up to 40°. Recorded dyke orientations show maxima in 143/74° NE and 98/approximately vertical. The former trend is parallel to the dominant trend of extensional faults in the region while the latter is parallel to the transfer faults in southern Svartenhuk Halvø. If it is assumed that deviation of the dykes from vertical is largely the result of tilting, then dykes trending parallel to the strike of the lavas have not in general been tilted as much as the lavas, implying that some of the tilt of the lavas took place before the latest volcanism in this area. However, around Tasiusaq several dykes with lithologies corresponding to the Svartenhuk Formation have been observed at right-angles to the dipping lavas, i.e. with north-east dips of 60–65° (Fig. 6), indicating that the dykes were tilted together with the lavas and that the entire tilting of the lavas here took place after intrusion of the dykes. However, some NE-dipping dykes appear to have been intruded into tilted fractures after tilting of the lavas. For example, one dyke has been observed dipping 45° NE and showing a concentration of cumulate mafic minerals along its lower contact, indicating that this dyke was intruded into a fissure that was already inclined before the dyke was intruded. Both at Kap Cranstown and east of Tartuusaq there are NE-dipping dykes that follow inclined fault planes and hence their dip need not be the result of later tilting. Further evidence that some NW–SE-trending dykes were intruded late in the history of the area is provided by the dykes cutting dolerite sheets on Itsaku; these are described below.

A proper study of the age relations between different dyke swarms and between dykes and faults was more than could be achieved with the resources allocated to the regional mapping project. Furthermore, trends observed in the early picrite and olivine-rich dykes were also utilised by younger dykes, so trend alone cannot be used to establish a complete dyke chronology. Intersections also show that dyke direction is no indication of relative age. For example, NE–SW-trending dykes have been observed both cutting and cut by dykes trending approximately 155°.

Some indication of the relative ages of dykes can be obtained from relationships between faults and dykes. For example NW–SE- and WNW–ESE-trending dykes, both of which follow the trends of major fault systems, may be strongly jointed along their margins, suggesting movements later than intrusion. Offset of both NE–SW- and approximately E–W-trending dykes at faults within the flexure zones (see p. 28) suggests that these dykes are older than development of the flexure zones, while other dykes are unaffected by these faults. There are also small right-lateral offsets of dykes on faults trending 100°, parallel to the transfer faults described in a later section. However, dykes are known to side-step on meeting planes of discordance, so this evidence is not conclusive. Within the flexure zones there are only very few NW–SE-trending dykes, i.e. dykes parallel to the zones.

Sills and sheets (Siuteqqut intrusions)

Extensive, thick (up to 150 m) sills and sheets of dolerite, and at least one sheet of picrite, occur along and in the vicinity of the Cretaceous boundary fault system on Svartenhuk Halvø and are best exposed within the Cretaceous–Palaeogene sediments on Itsaku; only the largest outcrops of these intrusions are shown on Fig. 2. For convenience they will be collectively referred to here as the Siuteqqut intrusions. In addition to the outcrops shown on Fig. 2, there are important sheets on the south-east side of Innerit, on the north-east side of Simiuttap Kuua, north of Qorlortup Kuua and south of Firefjeld. A sheet on the north-east side of Simiuttap Kuua is intruded along the inclined contact between basement and sediments or volcanic rocks before it transgresses through the Vaigat Formation. On entering the sediments between the Vaigat and Svartenhuk Formations it passes into a sill from which an irregular sheet has been fed into the overlying Svartenhuk Formation. The sheet on the south-east side of Innerit also reaches the intrabasaltic sediment horizon, but does not penetrate the lavas of the Svartenhuk Formation.

On Itsaku the very large representative of the Siuteqqut intrusions is cut by NW–SE-, NE–SW- and N–S-trending dykes. It is also cut by a significant NW–SE fault.

Situated as they are in the vicinity of the Cretaceous boundary fault system, the Siuteqqut intrusions are analogous to the Tartunaq intrusions on Nuussuaq which were intruded both along the plane of the Cretaceous boundary fault in Saqqaqdalen and into sediments and gneisses on either side of the fault (Munch 1945; Pulvertaft 1989). The Tartunaq intrusions have been dated at 54.8 ± 0.4 Ma (Storey *et al.* 1998), i.e. they are younger than the Vaigat and Maligât Formations on Nuussuaq, and they are also younger than the latest movements on the Cretaceous boundary fault system (Pulvertaft 1989). As just described, some of the Siuteqqut intrusions cut across the Vaigat Formation and locally penetrate into the Svartenhuk Formation, but outcrops of these intrusions are not good enough to allow any comment on the age of these intrusions relative to movements on the Cretaceous boundary fault system. In any case any deformation of the Siuteqqut intrusions along the boundary fault could be related to later inversion on this fault (see p. 22), a feature not observed on the Cretaceous boundary fault system on Nuussuaq.

If the Siuteqqut intrusions are not only younger than the Vaigat and Svartenhuk Formations but also younger than the main phases of faulting on Svartenhuk Halvø, the dykes cutting the Siuteqqut sheet on Itsaku are amongst the youngest intrusions in the area. The occurrence of dykes that are younger than both the basalts and the main fault movements would accord with what has been observed elsewhere in the West Greenland basalt province; for example in western Disko there are intrusions up to 25 Ma younger than the youngest lavas (Storey *et al.* 1998), and some of the lamprophyre dykes on Ubekendt Ejland are at least 18 Ma younger than the youngest known basalts (Parrott & Reynolds 1975; Storey *et al.* 1998).

Basin development and the Cretaceous boundary fault system

Most of the faulting seen on Svartenhuk Halvø took place after eruption of the Svartenhuk Formation lavas. However, the Cretaceous boundary fault system has a much longer and more complex history, as have also the faults on Itsaku.

Unlike the younger faults on Svartenhuk Halvø, the Cretaceous boundary fault system has an irregular course across the peninsula, just as it has in its continuation to the south-south-east (Fig. 1; Chalmers *et al.* 1998, 1999). Nothing is known about its course in the Innerit area north-west of Umiiarfik fjord. South-east of Umiiarfik the fault system runs south-east along the Simiuttap Kuua valley. In the inner part of this valley the fault turns abruptly to slightly north of east before turning again to follow a north–south valley until reaching the Qorlortup Kuua valley where it strikes NW–SE. In the outer part of the Qorlortup Kuua valley the position of the fault system is speculative because of the very extensive cover of fluvio-glacial sediments. It appears to split into four branches. One branch strikes approximately E–W and runs into the bay Kangiusap Imaa; this marks the northern boundary of outcrops of Cretaceous sediments and is the boundary fault proper. Two branches cross Itsaku, and another branch runs in 147° into Umiiviup Kangerlua. Not only do fluvio-glacial deposits conceal the fault system here but also extensive dolerite sheets (Siuteqqut intrusions) obscure the faults in this area. Cross-sections across the Cretaceous boundary fault system are shown in Plate 1.

Rifting in the Cretaceous basin probably started in Albian or earlier time (Chalmers *et al.* 1998, 1999), but it is not known whether the boundary fault system on Svartenhuk Halvø was initiated at this time or later. The Upper Albian – Lower Cenomanian deltaic deposits on the north side of Itsaku could be syn-rift sediments deposited by deltas prograding parallel to the boundary fault, but they could also represent part of a pre-rift delta complex that extended over the area north of the fault.

The Turonian – lower Campanian marine mudstones and distal turbidites were deposited in a completely different environment, far from fluvial influence. At this time the depositional area almost certainly extended north-east of the fault zone (see p. 10), and it is suggested that these sediments were deposited during a marine transgression that accompanied a phase of thermal subsidence.

Following Turonian–Campanian marine sedimentation there was a phase of uplift and erosion. If a parallel is drawn to developments on Nuussuaq (Dam & Sønderholm 1994; Dam *et al.* 1998a, 2000), this uplift could have taken place either during the Campanian or Maastrichtian, or perhaps not until early Paleocene. During this phase most if not all Cretaceous sediments were eroded off the basement area to the north-east, while conglomerates and other clastic sediments were deposited in erosional lows in the basin area (Fig. 7). Following this there was renewed uplift and erosion, so that any ?upper Campanian – Maastrichtian and lowermost Paleocene sediments that were present were removed everywhere except in the Itsaku block which must have remained a depressed fault block at this time. During this phase of erosion any surviving outliers of Cretaceous sediments in the basement area were also removed, and there was probably further removal of Upper Cretaceous sediments in the basin area.

In mid-Paleocene time the tectonic regime changed to one of subsidence of the basin area along the boundary fault system. Following erosion of the fault plane, mudstones and very coarse conglomerates were deposited and the foot of the bevelled fault scarp. Subsidence was followed rapidly by the eruption of contaminated and picritic basalts of the Vaigat Formation in a subaqueous environment. While hyaloclastite breccias encroached into the area from the south and south-west, mudstones and siltstones, the time-equivalents of the lower hyaloclastites to the south-west and west, were deposited close to the scarp. As the basin became filled up with basalts and subsidence waned, the basalt pile periodically emerged above water level, so that in the south-east there is an interval of alternating subaerial lavas and hyaloclastite breccias. In due course the hyaloclastites and subaerial lavas overstepped the boundary fault zone to lie directly on the basement to the north-east (Fig. 2). The uppermost basalts of the Vaigat Formation are entirely subaerial, with strongly reduced thicknesses in the north and on Innerit.

Evidence of continued tectonic activity prior to the eruption of the Svartenhuk Formation is provided by

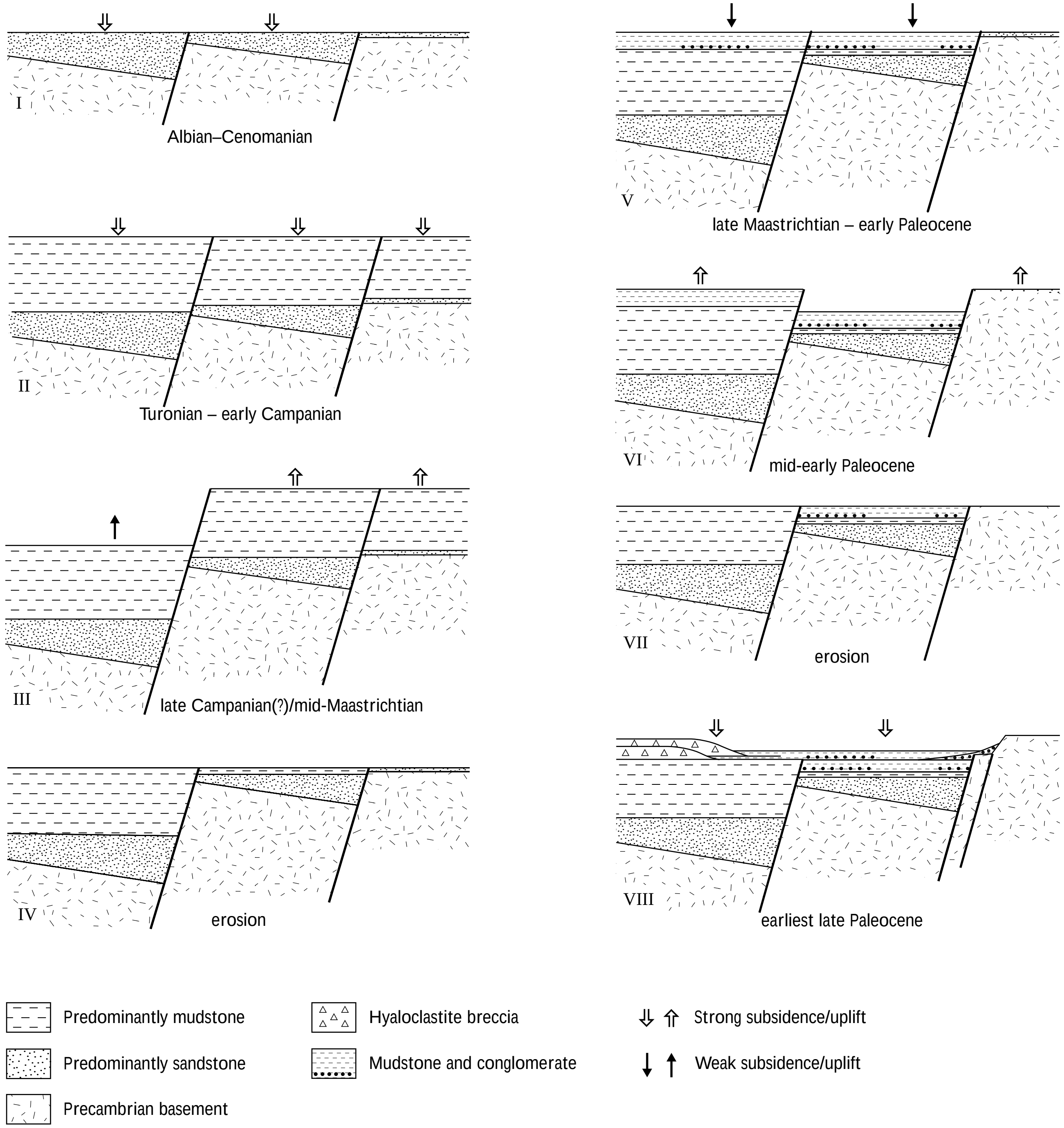

Fig. 7. Diagrammatic representation of the evolution of the Firefjeld–Itsaku area in Late Cretaceous – early Paleocene time.

the angular unconformity locally seen between lavas of the Svartenhuk and Vaigat Formations in the east central part of the peninsula. In the south-west, however, there was no significant break between eruption of Vaigat Formation and Svartenhuk Formation lavas. In contrast, northern Svartenhuk Halvø lay at the fringe of volcanic influence at this time, and here a fluvial plain with lakes developed which straddled the boundary fault and extended some way to the north. The earliest Svartenhuk Formation lavas here flowed into these lakes and developed as hyaloclastite breccias and pillow lava. Stable conditions followed as lavas of the Svartenhuk Formation spread over the basement far to the east and north-east.

The final movement along the Cretaceous boundary fault system on Svartenhuk Halvø took place at some time after the extrusion of the Svartenhuk Formation, and is the best documented; its net effect can

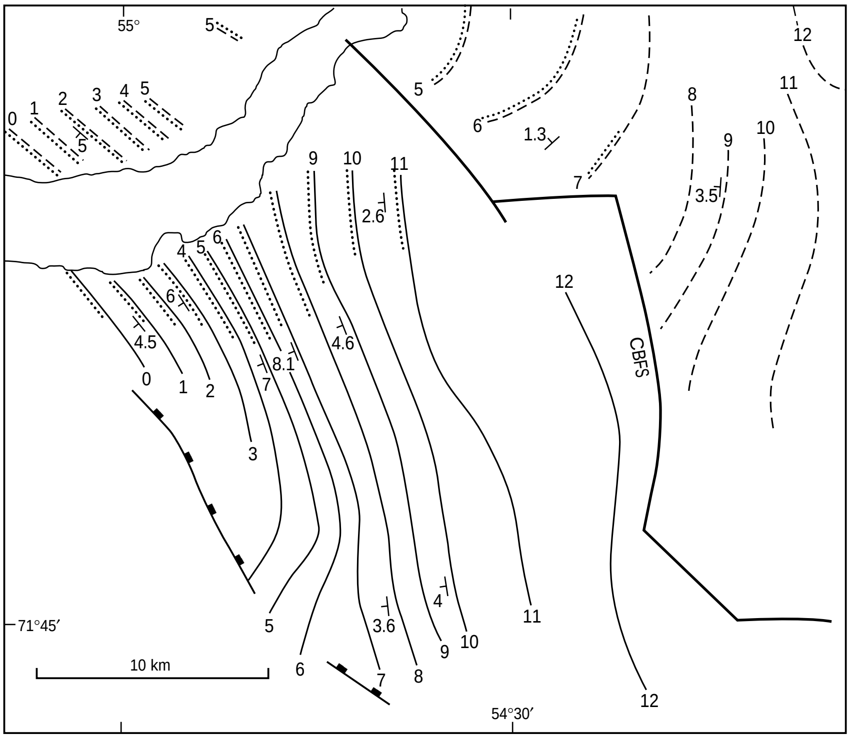

Fig. 8. Structural contour map of the base of Svartenhuk Formation in the north-western part of Svartenhuk Halvø. Contour interval 100 m. Dots along contours indicate where the Svartenhuk Formation overlies intrabasaltic sediments. **CBFS**: Cretaceous boundary fault system.

be seen in the structural contour map of the base of this formation (Fig. 8). This shows uplift of the basin area south-west of the boundary fault relative to the area to the north-east, i.e. *inversion* (cf. Rosenkrantz & Pulvertaft 1969; Münther 1973). Furthermore, the differential post-Svartenhuk Formation displacement increases from 200 m in the east to 600 m in the west (Fig. 8). The outer part of the Innerit area is also depressed relative to the area south-east of Umiiarfik, indicating the presence of a fault along this fjord.

Although evidence of inversion along the northern part of the Cretaceous boundary fault system is unequivocal, no definite signs of inversion have been observed in faults in the basin area. However, the stratigraphic control of complex fault displacements in the flexure zones south of Usuit Kuussuat is poor (see p. 27), and it is here in particular that further evidence for inversion should be sought during future field work in the area. Unfortunately, no information is available that could throw light on the timing and cause of the inversion. Inversion structures have been identified offshore west of Disko and in north-east Baffin Bay. Offshore west Disko open anticlines expressed by the top-basalt surface were attributed by Whittaker (1996) to transpression along a N–S-trending transfer fracture system (see p. 33). In the offshore sedimentary basins in north-east Baffin Bay sediments were folded into anticlines at about the same time as the Eurekan orogeny led to uplift of arches in the Canadian Arctic Islands (Whittaker *et al.* 1997). These examples, however, are far removed from Svartenhuk Halvø and there is as yet no evidence to justify relating inversion on Svartenhuk Halvø to either of them.

The maximum present-day height of hyaloclastite breccias recorded in the basin area is 950 m on Itsaku and 900 m on Firefjeld; these figures provide a measure of the minimum post-Paleocene uplift in this area. In the basement area to the north-east hyaloclastite breccias have been observed in the Svartenhuk Formation up to 1350 m above sea level; if these breccias were extruded in a marine and not a freshwater lake environment, post-Paleocene uplift of this area was at least of this order. This uplift is thought to have taken place in the Neogene (Japsen & Chalmers 2000; Chalmers 2000), although in inner Svartenhuk Halvø of the order of 100 m of the post-Paleocene uplift can be attributed to post-glacial (Holocene) isostatic rebound (Weidick 1976). No satisfactory mechanism has yet been proposed to account for the Neogene uplift that has taken place not only in West Greenland but also all around the North Atlantic (Japsen & Chalmers 2000).

Structures within the basin area

Dips

A striking and significant feature of Svartenhuk Halvø is the way in which it is divided into dip domains with contrasting values of dip (Fig. 9), whereas the strike is consistently between 135° and 155° throughout most of the area. Sailing along the south coast of the peninsula from Ulissat to Kap Cranstown one sees lavas with dips consistently to the south-west at angles up to 40°; these belong to the Tartuusaq dip domain. In contract, along the coast of Umiiarfik which is along strike from the steeply dipping lavas of the south coast, one sees lavas that for the most part dip south-west at angles less than 10°. In the extreme west of the peninsula the lavas are horizontal or dip at very low angles towards both north-east and west. North-east of a line extending roughly from Qooroq to Ulissat dips are generally low, less than 6°, and there are few faults. This line corresponds to the position of a series of flexure zones with offsets at transfer faults (see p. 25). North-east of the boundary fault system the lavas are virtually horizontal, although the contours of the base of the Svartenhuk Formation show a very gentle dip to the west.

As can be seen from the map Fig. 2, the strike of the lavas in the basin area is everywhere parallel or nearly parallel to the extensional faults and ‘flexure zones’ described in the following sections, the only deviations from the dominant 135–155° strike being westerly dips close to the few N–S extensional faults and anomalous southerly or south-easterly dips in the Vaigat Formation in an area south of Usuit Kuussuat. Thus it can be concluded that the dip of the lavas is due to tilting of fault blocks. It can also be read from Figs 2 and 9 that increase in dip from north-west to south-east takes place mainly at well-defined boundaries between dip domains and not by gradual twisting of the lava pile. The best-defined boundaries are either flexure zones, extensional faults or transfer faults (see p. 29).

Extensional faults

Extensional faults are a prominent feature of the basin area, apart from those parts of the area lying north-east of Qooroq and Qiterlikassak and in the extreme west where only minor faults have been observed. In contrast, Innerit and the basement area to the north-east and east of the Cretaceous boundary fault system have not been affected by faults of any significance (Henderson & Pulvertaft 1987). The most important fault of all in the area is the boundary fault system itself, the complex history of which has already been described.

Extensional faults occur in four main directions. By far the greatest number of faults trend between 145° and 150°. In eastern Svartenhuk Halvø there are several faults trending *c.* 125° and at Kap Cranstown there are extensional faults trending between 100° and 115°, while in the central part of the peninsula a few faults trend N–S. In southern and central Svartenhuk Halvø there are several faults and fractures trending *c.* 100°; these are faults with both lateral and vertical displacements, the latter usually being a downthrow to the north. These are transfer faults and are described in a later section (see p. 29).

Faults trending between 145° and 150° show a distinct concentration in zones; in the areas in between faults are fewer and also show less displacement. Most of these faults show downthrow to the north-east, but their dip is often difficult or impossible to estimate, especially in the case of faults running along valleys. Dips recorded are between 45° and 85° to north-east.

One prominent 145–150° zone extends across the peninsula from the coast of Umiiarfik between Nuuit and Qooroq in the north-west to Saviit in the south-east, with a left-lateral shift on the Qiterlikassak transfer fault (see p. 30). The degree of faulting increases south-eastwards. North and south of the Usuit Kuussuat valley the faults are easily recognised due to displacement of the base of the Svartenhuk Formation. Between Qiterlikassak and Usuit Kuussuat the fault blocks are tilted 13–16° towards south-west (Plate 1). The added vertical displacement of five faults across a 3.7 km wide zone here is *c.* 900 m; the net extension across this zone is not known accurately because of uncertainty concerning the dips of the fault, but must lie between 6 and 12%. Between Usuit Kuussuat and the Saviit–Qinerfik transfer fault (see p. 30) both the fault zone and the spacing between faults are wider, and tilting of fault blocks lessens; the largest vertical displacement recorded on a single fault here is 400 m down to north-east.

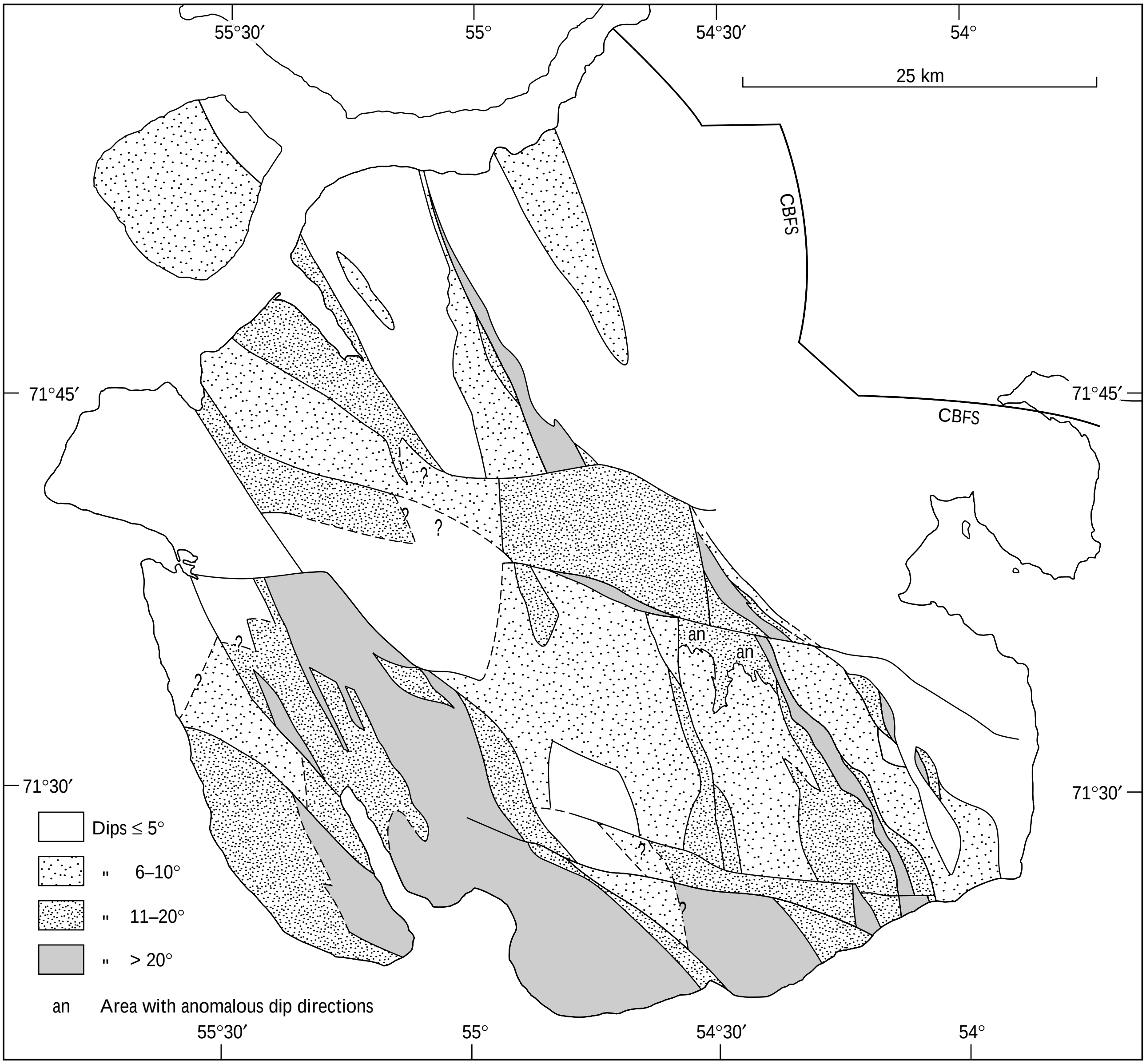

Fig. 9. Map of dip domains on Svartenhuk Halvø. Note that isolated, aberrant dips occur within some of the domains. **CBFS**: Cretaceous boundary fault system.

As has been discussed in an earlier section (p. 14), the apparent thickness of the Vaigat Formation in southern Svartenhuk Halvø is unlikely to be the true thickness. Stratal shingling, extensional faults with downthrow to the north-east, and a combination of these factors, could all account for this exaggerated apparent thickness of the formation. Several NW–SE-trending faults with downthrow to the north-east have been observed in southern Svartenhuk Halvø, but fault displacements can seldom be determined because the lack of distinctive marker horizons in the Vaigat Formation. *If* faults alone account for the excess thickness, the cumulative vertical displacement of these faults would have to be of the order of 4 km, and the cumulative extension *c.* 2 km or *c.* 15%. This question will be taken up in the final section of this paper (see p. 34).

An important fault zone trending between 145° and 150° runs north-westwards from Arfertuarsuk. The fault in Arfertuarsuk fjord itself has the largest displacement of any of the faults in this direction, the reappearance of the Vaigat Formation on the south-west side of the fjord and dip of the lavas requiring a downthrow of more than 2 km on the north-east side of the fault. The displacement of the Arfertuarsuk fault diminishes to the north-west, and the fault terminates at the E–W transfer fault through Svartenhavn. A fault running south-east from Milloofik has a maximum downthrow of 400 m to the north-east.

The overall structure seen between Maligissap Kuua and Tasiusaq can be described as an irregular, faulted roll-over into the Arfertuarsuk fault (Plate 1), assuming that this is a listric fault, and the nearby faults in 145–150° are synthetic faults with regard to this structure. This is also how the Arfertuarsuk fault has been described by Geoffroy *et al.* (1999). Without accurate information on the dip of the Arfertuarsuk fault, one cannot calculate where the fault levels out into a detachment zone, but it is suggested that that the detachment could be at or near the base of the sediments that underlie the basalts.

A suite of faults trending 146° crosses Skalø and continues into the mainland to the south-east. Displacements on the faults in this zone are relatively small, less than 150 m, with one exception that has a downthrow of 225 m to north-east.

Extensional faults trending *c.* 125° occur in eastern Svartenhuk Halvø between Umiiviup Kangerlua and Maniiseqqut. Most of these faults downthrow to the north, the largest downthrow on a single fault being 160 m. These faults were active during eruption of hyaloclastites in the Vaigat Formation, giving rise to the fault scarps already described. Some of these faults are splays at the termination of the strike-slip/transfer fault in the Usuit Kuussuat valley (see p. 31).

The faults at Kap Cranstown with trends between 100° and 115° all have downthrow to the north. The largest of these faults has a vertical displacement of 360 m.

North–south faults have only been recorded in the central part of the peninsula. Downthrows both to the east and to the west have been observed. In the vicinity of these faults the lavas strike N–S, which is abnormal for the area as a whole.

‘Flexure zones’; narrow half-grabens

A number of zones with abnormally steep dips cross Svartenhuk Halvø in directions between 133° and 154°, i.e. approximately parallel to the main extensional faults. Superficially many of these zones resemble monoclines, in that when crossing the zones from north-east to south-west one passes from an area with dips less than 12° to south-west through a zone with much steeper south-westerly dips and back into an area with south-westerly dips lower than 15°. These zones have therefore been referred to as flexure zones, but on closer examination it can be seen that they are really a special type of fault zone. The term ‘flexure zone’ is nevertheless retained to distinguish these zones from other structures in the area. In most cases, unfortunately, the displacement on the faults in the flexure zones is not known due to lack of stratigraphic control, and furthermore the dips of the fault planes are seldom evident and even the dip direction of faults may be uncertain.

The simplest flexure zone and that most resembling a monocline is that at Nuuit. Here, passing south-westwards, the south-westerly dip of the basalts increases abruptly from 3–6° to more than 15°. The change takes place at a nick-point without any apparent faulting. To the south-west one crosses a number of faults with downthrow to the north-east before the dip levels out to less than 10°.

The ‘flexure zone’ in the Qooroq valley is the only one where displacements on some of the related faults are known (Fig. 10a). The increase in south-westerly dip here takes place abruptly at a fault. Where the cross-section has been drawn the flexure zone is *c.* 2 km wide. The net displacement across the flexure zone here is a drop of the south-west side of about 600 m relative to the north-east side. The zone of steep dips, however, is downthrown in relation to both sides of the zone, i.e. the zone is a narrow half-graben. The steeply dipping basalts have been displaced by at least four small faults, each with downthrow to the north-east. While the north-east dips of the minor faults within the graben are distinct, it has not been possible to measure the dip of the major faults, although the impression is that they are steep.

The Qooroq flexure zone is also seen at Oqaasaq where, however, the south-western side of the zone is concealed by fluvio-glacial deposits. At Oqaasaq the net down-to-south-west displacement caused by faulting and flexuring across the zone is estimated to be 550 m. The abrupt steepening of the south-westerly dip of the basalts takes place at a steep fault. This fault is offset 600 m left-laterally at a fault trending almost N–S, with dip and downthrow to the east; a dyke has been emplaced along this fault. Within the flexure zone the basalts dip up to 26° to south-west and have been displaced by numerous faults, each with dip and downthrow to the north-east. Within the wedge between the major NW–SE fault and the *c.* N–S fault the strike of the tilted fault blocks is intermediate between that of the major NW–SE fault and the N–S fault, suggest-

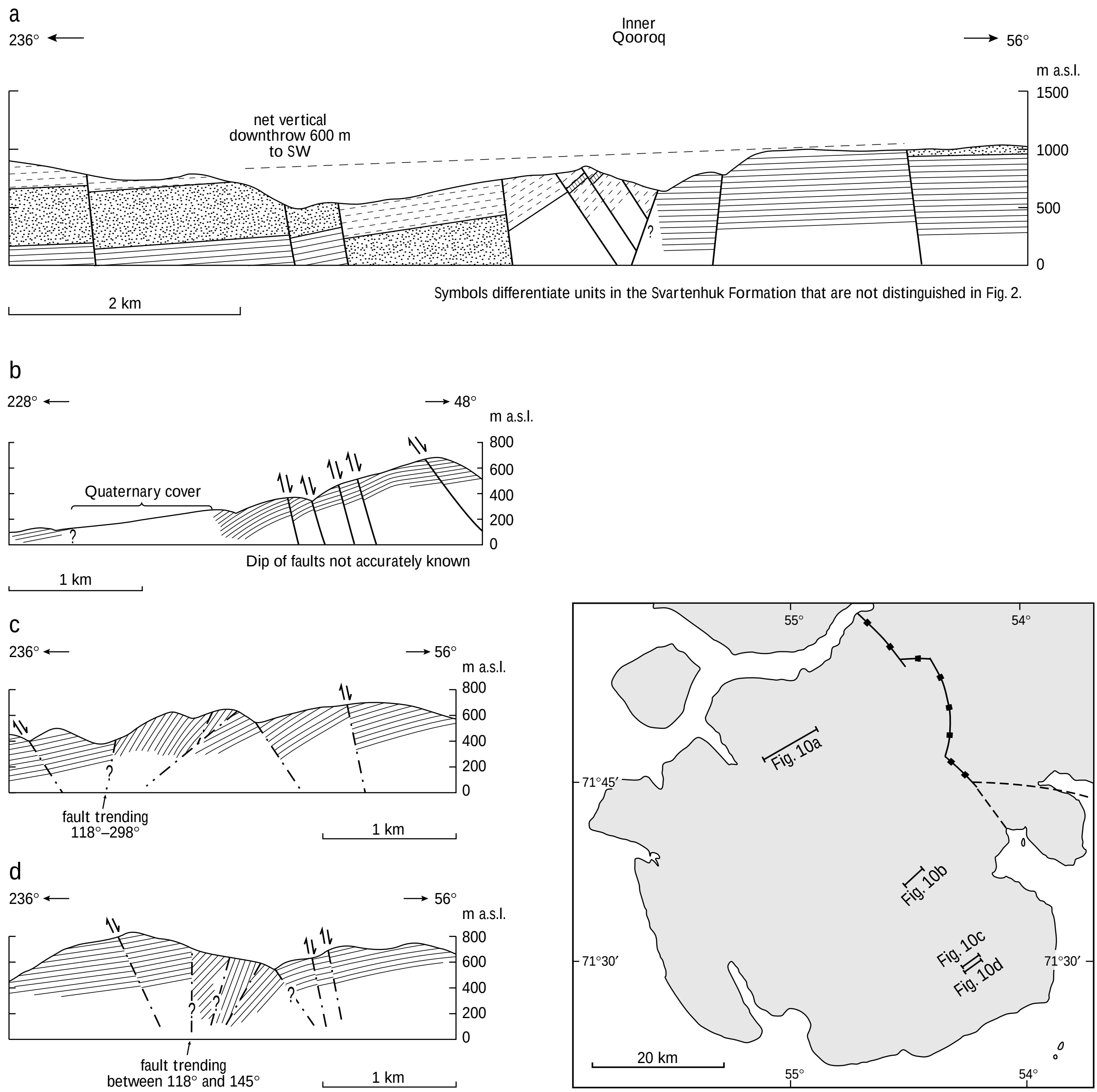

Fig. 10. Cross-sections of 'flexure zones' on Svartenhuk Halvø. **a**: Qooroq; **b**: between Qiterlikassak and Usuit Kuussuat; **c** and **d**: mountains between Usuit Kuussuat and Ulissat. For further explanation, see text.

ing that these two faults were active simultaneously. The transition from steep to gentle south-westerly dips on the south-west side of the zone is not exposed, but it is expected that it takes place at a fault. Close to the expected position of this fault the basalts locally dip 40° towards south-west.

The flexure zone between Qiterlikassak and Usuit Kuussuat has some of the characteristics of a true monocline in that the increase in dip from north-east to south-west starts gradually as a genuine flexure (Fig. 10b). However, as the dip increases, the basalts become displaced and tilted by several small faults with dip and downthrow to the north-east. South-westerly dips as steep as 43° have been recorded in this zone. Unfortunately, as at Oqaasaq, due to poor exposure it cannot be seen how the change from the zone of steep

Fig. 11. South-west margin of flexure zone 10 km south of Usuit Kuussuat, viewed from the south-east. The width of the area shown is *c.* 350 m. For location, see Fig. 2.

dips to the area with lower dips to the south-west takes place.

The most spectacular and complex flexure zone is that running from the Usuit Kuussuat valley to the south coast of Svartenhuk Halvø between Ulissat and Saviit. This zone is entirely within rather monotonous lavas of the Vaigat Formation in which there are very few marker horizons, so rarely can one observe the displacement on the faults in the flexure zone. Furthermore, it has not been possible to demonstrate a net relative down-to-south-west displacement across this flexure zone as could be recorded along the Qooroq–Oqaasaq flexure zone.

Ten kilometres south of Usuit Kuussuat the south-west flank of the flexure zone is well exposed at one locality. This is shown in Fig. 11. The fault itself is vertical or dips steeply to the north-east. Between this fault and the fault to the north-east (right) there is a wedge of near-vertical flows, while the lavas on the north-east side of the faults dip up to 58° to the south-west. It seems that a set of flows outcropping at the top of this wedge has been downthrown about 90 m to the south-west on the vertical fault. Several other extensional faults occur north-east of this and some can be seen in the background in Figs 11 and 12. These faults all dip and have obvious downthrow to the north-

Fig. 12. Panorama of the flexure zone in the mountains between Usuit Kuussuat and Ulissat, seen from the south-east; the cross-section Fig. 10d illustrates the same part of the flexure zone and gives the scale. Note the dyke standing out as a ridge and trending 40° that crosses the flexure zone with small right-lateral offsets at faults. For location, see Fig. 2.

east; the basalts between these faults have been tilted up as much as 48° to the south-west.

Figure 10c and d are cross-sections of the Ulissat – Usuit Kuussuat flexure zone 9.5 km north-west of Ulissat, and Fig. 12 shows the same part of the flexure zone as shown in Fig. 10d. Approaching from the north-east the increase in south-westerly dip starts at an extensional fault with dip and small downthrow to the north-east. Where the dip steepens abruptly, more faults are found. Finally, the dip steepens to more than 45° to south-west and at one locality is steeper than one of the faults. On its south-west side the flexure ends abruptly at a steep oblique fault trending 118°, south-west of which dips in the lavas are much lower than within the flexure zone. A lesser fault with dip and downthrow to the north-east occurs a short distance to the south-west (Fig. 10c, d).

The behaviour of dykes in the Ulissat – Usuit Kuussuat flexure zone is puzzling. One dyke with trend 40° and steep south-easterly dip shows right-lateral offset at four faults in the flexure zone (Fig. 12); the cumulative right-lateral offset is 200 m. In contrast, vertical dykes trending E–W or WNW–ESE show small left-lateral offsets at faults in the flexure zone. Other dykes cross the flexure zone without any offset. As already pointed out, offset of a dyke at a fault as observed on aerial photographs is not proof of fault movement later than dyke emplacement, as dykes often side-step at discordant surfaces; observations made on the ground are needed in this area. It may be significant that within the flexure zones there are only very few dykes parallel to the zones.

On the south coast the Ulissat – Usuit Kuussuat flexure zone is about 2¼ km wide. When viewed from the sea the impression gained is that of increasing south-westwards dip of the basalts taking place stepwise at extensional faults with dip and downthrow to the north-east. However, this impression is an oversimplification because faulting associated with the WNW–ESE-trending Saviit–Qinerfik transfer zone (see p. 30) interferes with the flexure zone here.

Interpretation of the mechanism(s) that gave rise to the flexure zones on Svartenhuk Halvø is hampered by the fact that only limited quantitative information is available concerning displacement and dip of the associated faults. We do know, however, that the Qooroq–Oqaasaq flexure zone is associated with a net relative downthrow of the block to the south-west. We also know that the basalts involved in these structures form a competent layered formation that overlies a substantial thickness of sediments dominated at least in their upper part by incompetent mudstones, and that these sediments lie on a massive, brittle basement. Furthermore, we know from Nuussuaq to the south that extensional faults were established in the basement before eruption of the basalts (Chalmers *et al.* 1999). With this in mind we suggest two possible mechanisms that could have given rise to the flexure zones and associated faults.

The most likely explanation is that the typical narrow flexure zones were developed above faults in the underlying basement that were reactivated after eruption of the basalts. In the situation on Svartenhuk Halvø where an incompetent formation separates the competent basalts from the massive basement, the resulting structures in the basalt cover will not be a simple upwards projection of the basement fault. On the contrary, results of model experiments involving reactivation of a fault in a massive 'basement' overlain by an incompetent layer that is overlain in turn by a competent cover have shown that reactivation of a fault in the basement can lead to a variety of structures in the cover including monoclinal folds, grabens and even reverse faults and thrusts (Mandl 1988, fig. I.2-37; Oudmayer & de Jager 1993, figs 7, 8; Vendeville *et al.* 1995, fig. 6). Pascoe *et al.* (1999) have forward-modelled the structures generated in experiments by Vendeville *et al.* (1995), producing the result shown in Fig. 13a. This structure shows a striking resemblance to the structure in the Qooroq flexure zone/half-graben shown in Fig. 10a, but it should be borne in mind that the relative thickness of the incompetent layer (i.e. mudstones) underlying the basalts at Qooroq is probably much greater than the incompetent layer in Pascoe *et al.*'s forward model. Furthermore it is not known whether the structures on Svartenhuk Halvø ever became overlain by syn- and post-rift sediments as shown in Pascoe *et al.*'s forward modelling.

An alternative explanation is suggested by analogue models of extensional fault geometries reported by Cloos (1968) and McClay & Ellis (1987) and could apply to the flexure zones south-east of Qiterlikassak. The asymmetric graben produced experimentally by Cloos (1968) shows features in common with the flexure zone between Qiterlikassak and Usuit Kuussuat, in particular the way in which steepening of dip of layers in the flexure is accompanied by development of numerous minor faults dipping in the opposite direction (Fig. 13b). Fig. 13c shows the structure generated by McClay & Ellis above a listric detachment fault where the listric fault is almost vertical near the surface. Extension on this fault resulted in a well developed faulted

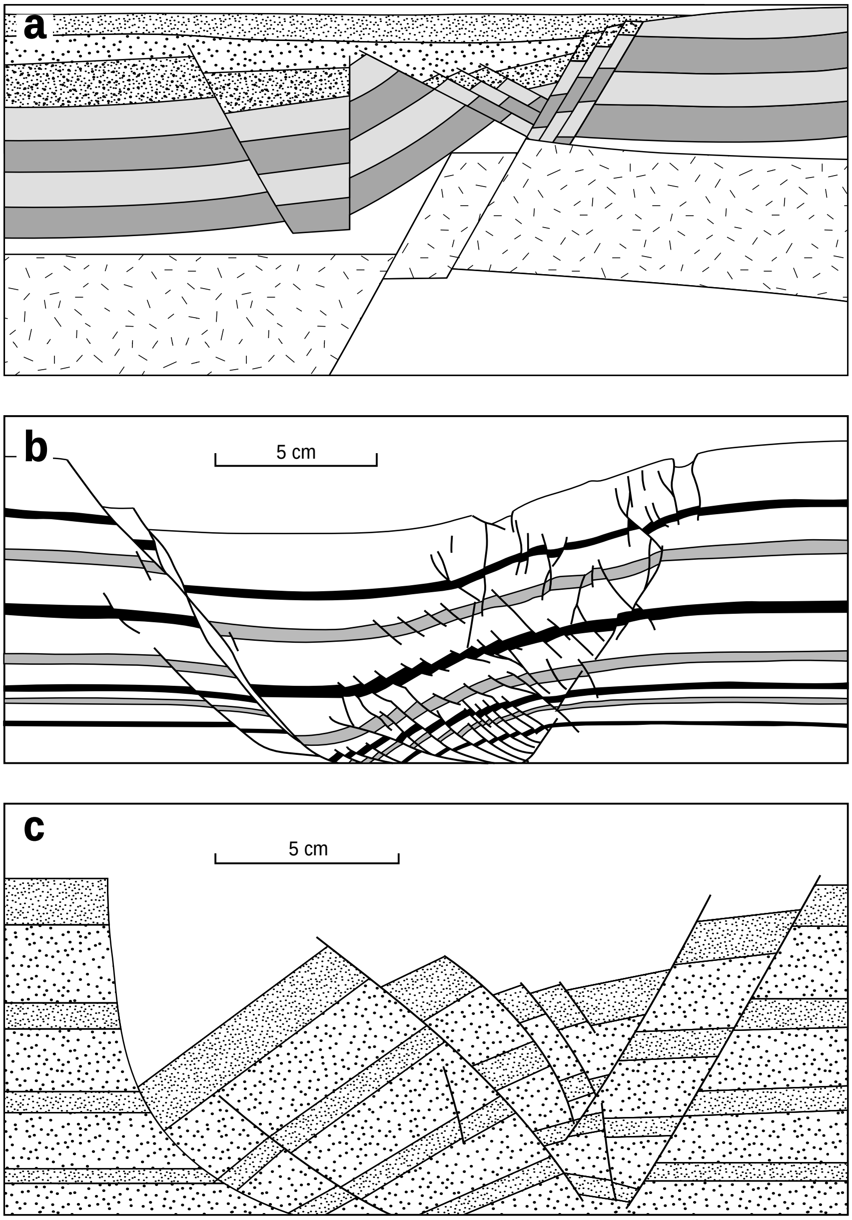

Fig. 13. Examples of fault and flexure patterns produced in analogue model experiments and forward models. **a**: forward model of the structure developed by reactivation of a basement fault where a competent cover is separated from the basement by a ductile layer (Pascoe *et al.* 1999, fig. 6d); **b**: asymmetric graben produced in a clay model (drawn from Cloos 1968, fig. 16, in mirror image); **c**: structure formed by 33% extension on a listric fault (McClay & Ellis 1987, fig. 5b in mirror image).

roll-over with highly rotated (up to 35°) layering; the steepest dip is seen in the segment closest to the listric fault. This modelled structure suggests that the flexure zone south of Usuit Kuussuat might have formed above listric faults that levelled out in detachment zones in the sediments below.

Transfer faults; strike-slip faults

In the southern and central parts of Svartenhuk Halvø there are several faults and fractures trending between 90° and 115°. The four most important of these are the Saviit–Qinerfik fault, the Svartenhavn fault, the Usuit Kuussuat fault and the Qiterlikassak fault. It is tempting to connect the Svartenhavn and Usuit Kuussuat faults, but we have no evidence for movements in the inner part of the Maligissap Kuua valley, nor has any significant change in dip been detected across this part of the valley as there is across the Svartenhavn, Qiterlikassak and Saviit–Qinerfik faults.

The Saviit–Qinerfik, Svartenhavn and Qiterlikassak faults are typical transfer faults as defined by Gibbs (1984, 1989) and Lister *et al.* (1986). Transfer faults allow extension to be transferred laterally along the strike of the fault and divide extensional terrains into segments. Characteristically the degree of extension changes abruptly across transfer faults, and in many cases – but not on Svartenhuk Halvø – the polarity of extensional faults also changes across transfer faults. Transfer faults are analogous to transform faults in spreading oceans, in that the slip on transfer faults is opposite to the sense of mapped offset of the zone of extensional faulting.

On Svartenhuk Halvø the sense of mapped offset of dip domains and extensional zones on the Qiterlikassak, Svartenhavn and Saviit–Qinerfik faults is left-lateral, whereas there is evidence of right-lateral movement on these faults.

The Svartenhavn fault is entirely concealed by alluvium or the sea. Its presence is indicated by an abrupt change in dip values from very low on the north side of the fault to more than 20° south of the fault. With a left-lateral offset this domain of relatively steep dips can be predicted to reappear on the north side of the fault in the offshore area. North of the fault there is an outcrop of the Arfertuarsuk trachyte that is displaced dextrally relative to the outcrops south of the fault. This can be taken as evidence of right-lateral displacement along the Svartenhavn fault, but it could also be the consequence of purely vertical displacements along the Svartenhavn and Milloofik faults. The lavas here have low dips, so the vertical displacement required to cause the repetition is not large.

The Saviit–Qinerfik fault extends from Saviit to Qinerfik, north of Tasiusaq. The fault separates a dip domain to the north with south-westerly dips less than 16° from the Tartuusaq dip domain to the south where dips generally are more than 25° and locally as much as 40° to south-west. Along this fault the dip domain boundary is offset 17 km in a left-lateral direction, whereas outcrops of the Kakilisaat Member at Ulissat are separated right-laterally by 10 km from the large outcrop of this member west of Saviit. If there were no other faults in the area, this separation would imply either a 10 km strike-slip displacement on the Saviit–Qinerfik fault or a *c.* 3 km downthrow to the north or oblique slip combining substantial lateral and vertical movements. However, a lateral displacement of 10 km is not plausible. At Qinerfik, a mere 25 km west of Saviit, the lateral displacement on the fault is at most 250 m. A 10 km lateral displacement could hardly be reduced to a few hundred metres within 25 km. Some right-lateral slip, however, is required, and in the southern part of the area there are examples of smaller faults trending *c.* 100° that displace vertical dykes in a right-lateral direction, proving that such displacements have taken place in this part of the area. There are at least two possible explanations of the situation described.

One explanation is that a new phase of eruption of lavas with Kakilisaat Member chemistry gave rise to the contaminated lavas west of Saviit, so that these form a younger unit than the Kakilisaat Member. This explanation is not favoured. As already mentioned, over a large area of south-east Svartenhuk Halvø, both north and south of the Saviit–Qinerfik fault, the contaminated lavas are overlain by a characteristic massive picrite flow with large olivine phenocrysts. This suggests that all contaminated lavas in this area belong to the same stratigraphic unit – the Kakilisaat Member.

The other explanation of the large apparent lateral displacement of the Kakilisaat Member is that faults trending between 145° and 150° have caused a repeated, stepwise relative uplift of the Kakilisaat Member to the south-west, without bringing it to the surface, so that north of the Saviit–Qinerfik fault this member is not far below sea level, while south of the fault its repetition is seen just west of Saviit. If this is the case, only a modest lateral slip on the Saviit–Qinerfik fault accompanied by a downthrow to the north is required to account for the present-day map pattern.

Other ESE faults and fractures occur in a 1.5–3 km wide zone north of the Saviit–Qinerfik fault. West of Ulissat some of these faults can be seen to have downthrow to the north or equivalent right lateral displacements. Dykes occur both within and parallel to these faults. These structures may be the result of transtension along the Saviit–Qinerfik fault.

The eastwards extension of the Saviit–Qinerfik fault under Karrat Fjord is not known. Chalmers *et al.* (1998) have shown it running south of Schade Øer, because these islands consist of rocks belonging to the Kakilisaat Member which can be linked to the outcrops at Maniiseqqut without introducing lateral displacements.

As already pointed out, the Saviit–Qinerfik fault dies out westwards at Qinerfik where it links into the extensional fault system. The Tartuusaq dip domain continues north-west to the Svartenhavn transfer fault.

Along the Qiterlikassak fault the sense of mapped offset is left-lateral, but no evidence of right-lateral slip on this fault has been recorded. In the valley here there are no proper exposures, but there are ridges in the terrain that run parallel to the fault. These appear to reflect the strike of steeply-dipping lavas under the Quaternary debris. The existence of a narrow zone of steeply-dipping lavas striking parallel to the fault can be explained by transtension along the fault.

In the Usuit Kuussuat valley the sense of mapped offset of flexure zones appears to be right-lateral, although a correlation of flexures zones involving a left-lateral offset across the valley could also be made. The existence of a right-lateral fault in this valley is strongly suggested by a small downthrown lens of steeply-dipping Svartenhuk Formation lavas in the inner part of the valley. The shape and orientation of this lens is like that of a small pull-apart basin resulting from transtensional displacement on a right-lateral fault trending 105° along the valley floor.

The Usuit Kuussuat fault dies out at both ends in extensional horsetail splays, best seen at the east-south-east end where the splay faults show downthrow to the north-east.

Origin of the structural pattern

It remains to be discussed whether and how dip directions and values, extensional faults, flexure zones, and transfer and strike-slip faults are related to one another and to the regional structural development and plate-tectonic setting of central West Greenland.

The Cretaceous boundary fault system is a segmented extensional fault system that extends from Svartenhuk Halvø in the north to the south-east corner of Disko Bugt in the south (Fig. 1). The fault system has been described in several papers, most recently by Chalmers *et al.* (1999) who interpreted its zig-zag course as partially controlled by pre-existing fractures in the Precambrian basement and also drew attention to its similarity to the course of the Suez rift (Patton *et al.* 1994). The fault appears to have been active in the Late Albian – Early Cenomanian on Qeqertarsuaq and Upernivik Ø (Chalmers *et al.* 1999), whereas on Nuussuaq the fault system truncates Albian–Cenomanian sediments without any sign that it was active during sedimentation here. Inversion on the Cretaceous boundary fault system has only been observed on Svartenhuk Halvø. Otherwise outcrops on Svartenhuk Halvø add little to what is already known from areas to the south-south-east.

As has already been pointed out, dip directions in the basalts in the basin area are predominantly normal to the strike of extensional faults. Most of these faults strike between 145° and 150°, and dip is largely the consequence of rotation along faults with this trend. The flexure zones are also interpreted as extensional structures that most likely developed as a response to reactivation of faults in the underlying basement with uplift on the north-east side. Alternatively they could have formed in connection with listric faults that levelled out below the base of the basalts. The question to be addressed now is whether the direction of *regional* extension is normal to the numerous extensional faults in 145–150°, because McClay & White (1995) have shown that this need not necessarily be so. In situations where extension in a rift system is oblique to the overall trend of the underlying rift, extensional faults within the rift cover can form parallel or at a low angle to the rift margins rather than at right angles to the direction of extension, even where the direction of extension is at 45° to the rift (McClay & White 1995, fig. 5D). Thus, if the rift system in the sediments and basement underlying Svartenhuk Halvø trends 145–150°, extensional faults in this direction could form in the basalt cover as a consequence of regional extension in any direction between *c.* 55–60° (i.e. normal to the extensional faults) and 100–105°. Now the latter direction, 100–105°, is approximately the trend of transfer faults on Svartenhuk Halvø. Transfer faults are by definition part of the extensional fault system (Gibbs 1984), and ideally they are parallel or nearly parallel to the extension vector. Thus one could suggest that on Svartenhuk Halvø the post-basalt regional extension vector was in 100–105° and that this acted on a deeper NW–SE rift system, giving rise to widespread extensional faults in 145–150°.

Even though approximately E–W extension could account for the extensional structures seen in the basalts on Svartenhuk Halvø and is consistent with the direction of transfer faults in the region, this suggestion is not favoured because it is not compatible with current models for the regional plate-tectonic setting of the region and the – admittedly sparse – information available from adjacent offshore areas.

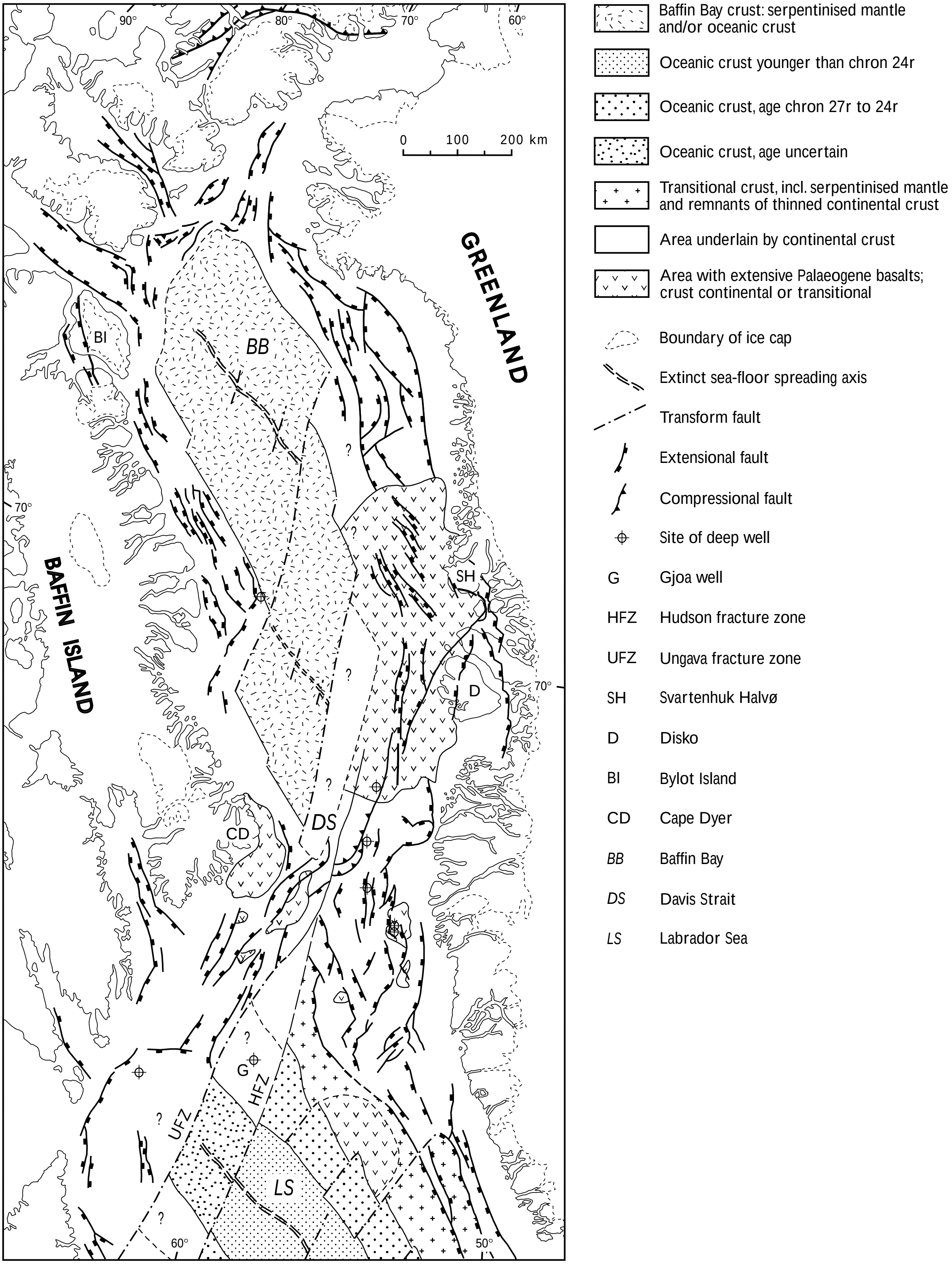

Fig. 14. Map of crustal structure of the northern Labrador Sea – Davis Strait – Baffin Bay region. Principal sources: Keen *et al.* (1974), Balkwill (1987), Jackson *et al.* (1992), Chalmers *et al.* (1995), Whittaker (1996), Reid & Jackson (1997) and Whittaker *et al.* (1997). Compilation by T.C.R. Pulvertaft.

At the time when syn- and post-basalt faulting was taking place on Svartenhuk Halvø, sea-floor spreading was active in the inner Labrador Sea. Spreading in Baffin Bay probably began at the same time or shortly after spreading started in the Labrador Sea, i.e. in the late Paleocene (Chalmers & Pulvertaft in press). Spreading in both regions slowed down after magnetochron 20 (43 Ma; middle Eocene) and had died out altogether by magnetochron 13 (33 Ma; earliest Oligocene) (Srivastava 1978; Chalmers & Pulvertaft in press). The spreading axes in the respective regions were linked by an approximately north–south transform fracture system (Fig. 14), and the area offshore west Disko was adjacent to this active transform system from late Paleocene until the end of the Eocene. Open folding of the top-basalt surface offshore west Disko has been interpreted as due to transpression associated with this transform system, and the N–S-trending extensional faults mapped here from the seismic data could be transtensional rather than simple extensional features (Whittaker 1995, 1996).

At about 70°30′N offshore, and on Ubekendt Ejland onshore, the structural trend changes fairly rapidly from *c.* N–S to NW–SE (Fig. 14). This is approximately where the current model (Chalmers & Pulvertaft in press) requires that the strike-slip influenced area gives way northwards to an extensional regime in which post-basalt faults are parallel to the extinct spreading axis in Baffin Bay. Svartenhuk Halvø belongs to this regime, and the NW–SE-trending extensional faults onshore and offshore are most likely controlled by regional extension normal to the faults and the spreading axis.

Accepting that, in accordance with the regional model, the direction of extension on Svartenhuk Halvø was *c.* 55–60°, the transfer faults on the peninsula are at about 40° oblique to the extension direction. Oblique transfer faults are known from many rift basins, e.g. the Midland Valley of Scotland and the Carlisle Basin, UK (Gibbs 1989), the Suez Rift (Chénet *et al.* 1987; Colletta *et al.* 1988; Moustafa 1997; McClay & Khalil 1998), the East African Rift (Ebinger 1989; Chorowicz & Sorlein 1992), the eastern margin of the Basin and Range province, USA (Henry 1998), the Rocôncavo Graben, Brazil (Milani & Davison 1988), just as transform faults can be oblique to the direction of extension in a spreading oceanic ridge, e.g. the Grimsey Fault in northern Iceland which is at an angle of about 40° to the extensional faults and grabens on either side (Gudmundsson *et al.* 1993). However, if transfer faults are markedly oblique to the direction of extension, there must be an element of extension or compression along the transfer (Gibbs 1989). On Svartenhuk Halvø there is evidence of transtension along the Qiterlikassak and Usuit Kuussuat faults, and there are extensional faults parallel to and associated with the Saviit–Qinerfik fault that can also be attributed to transtension. The numerous dykes parallel to this fault indicate that extensional fractures in this direction existed already during volcanism. However, no evidence of extension has been observed along the Svartenhavn fault, although just north-west of Svartenhavn there are a few dykes parallel to this fault. However, exposures in the Maligissap Kuua valley are very poor and such evidence could have been overlooked.

Of the examples of oblique transfer faults listed in the foregoing paragraph, that described by McClay & Khalil (1998) from the Araba – Abu Durba area on the eastern margin of the Gulf of Suez shows the closest similarity to the structural pattern on Svartenhuk Halvø. To make this similarity clear, a rotated mirror image of McClay & Khalil's map is shown in Fig. 15 side-by-side with a simplified map of the faults on Svartenhuk Halvø. Elements of similarity include the approximate angle between regional extension and the transfer faults, vertical displacement on the transfer faults, horsetail splays at the termination of transfer faults, and the way in which the Durba transfer fault of McClay & Khalil (1998) and the Saviit–Qinerfik transfer fault on Svartenhuk Halvø link into the extensional fault systems in the respective areas.

In rift systems where oblique transfer faults occur, the orientation of these faults is often controlled by older structures in the underlying basement (e.g. Patton *et al.* 1994; Ring 1994; Moustafa 1997). Although no E–W or ESE faults and shear zones have been recorded in the Precambrian basement east of Svartenhuk Halvø (Henderson & Pulvertaft 1987), there are WNW–ESE fractures in the basement north-west of Kangiusap Imaa, and E–W segments of the Cretaceous boundary fault system occur both north of Itsaku and in south-west Upernivik Ø (Figs 1, 2), so weaknesses in the basement in this direction do exist.

For completeness it should be mentioned that if the transfer faults were right-lateral strike-slip faults, the right-lateral couple would generate extensional faults in the direction of the dominant extensional faults on Svartenhuk Halvø. This is believed to be a coincidence and insufficient to counter the evidence against the E–W to 105°–285° faults being major strike-slip faults.

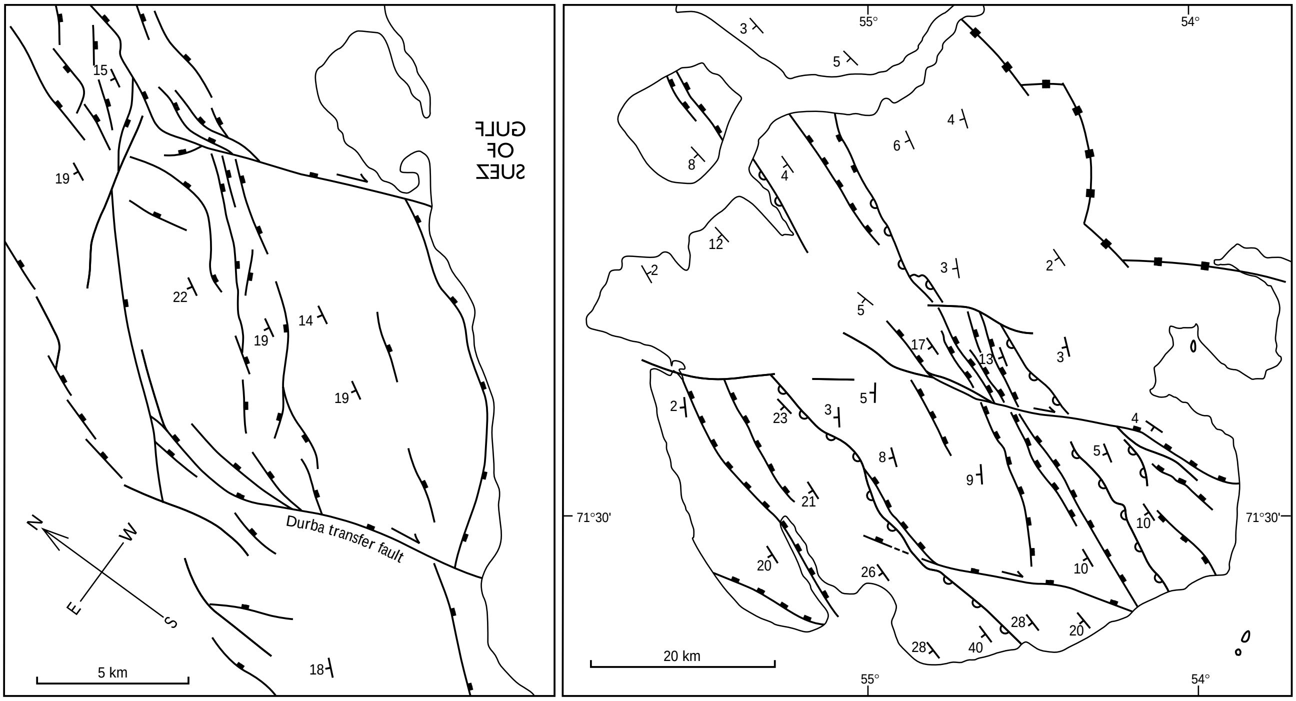

Fig. 15. Simplified, rotated **mirror image** of the map of the Durba and Ekma transfer faults, Araba–Abu Durba area, eastern Gulf of Suez (modified from McClay & Khalil 1998, fig. 1), side-by-side with a simplified map of Svartenhuk Halvø.

The relative roles of extensional faulting, stratal shingling and syn-volcanic tilting in southern Svartenhuk Halvø

It has already been pointed out that the very large apparent thickness of Vaigat Formation lavas in southern Svartenhuk Halvø (*c.* 9 km) presents a problem that has been discussed since early reconnaissance investigations (Noe-Nygaard 1942; Pulvertaft & Clarke 1966; Münther 1973; Larsen 1981b). While repetition by extensional faulting could alone account for this apparent thickness, the number of NW–SE faults actually observed along the south coast of Svartenhuk Halvø seems insufficient to be the only factor involved. However, significant faults may be concealed under the large valleys, and furthermore both the rubbly nature of outcrops in the Vaigat Formation and the uniformity of the formation make faults within the formation hard to recognise.

Stratal shingling (Crowell 1982) is also a feasible mechanism for the development of a lava succession with a stratigraphic thickness that greatly exceeds the real thickness of the succession. However, interpreting the excessive stratigraphic thickness of the Vaigat Formation in southern Svartenhuk Halvø as due to stratal shingling alone meets with difficulties. Essential to the mechanism is a basinward migration of the source accompanied by subsidence of the basin during volcanism, so that the younger lavas are confined to the basinward side of the outcrop. Retreat of lava flows from northern Svartenhuk Halvø did take place at the end of Vaigat Formation time to allow the deposition of the intrabasaltic sediments here, but this was succeeded by eruption of the Svartenhuk Formation which extends farther to the east and north-east than any of the older lavas, disappearing finally under the Inland Ice (Fig. 1). Furthermore, young dykes with lithologies corresponding to the lavas of the Svartenhuk Formation are very numerous throughout the Vaigat Formation in eastern Svartenhuk Halvø, which does not suggest any basinward (south-westerly) retreat of volcanic sources at this time.

Although stratal shingling *sensu stricto* does not appear to have been a major factor in the structural development of Svartenhuk Halvø, some subsidence did take place during eruption of the Vaigat Formation and was accompanied by flexuring and development of down-to-north-east extensional faults. Geoffroy *et al.* (1998) have provided structural data indicating that on Disko island flexuring and down-to-east extensional faulting was syn-volcanic, and the same team (Geoffroy *et al.* 1997, 1999) propose that this is also the case in southern Svartenhuk Halvø and on Ubekendt Ejland. They interpret the overall structure of Svartenhuk Halvø as a syn-volcanic flexure with fan geometry in cross-section, implying that the basalts on Svartenhuk Halvø were erupted into an active half-graben bounded to the south-west by the Arfertuarsuk listric fault. Geoffroy *et al.* (1997) have also observed a progressive seaward-retreating development of unconformities within the pile of lava flows. Our observations also suggest that some of the tilting of the Vaigat Formation lavas was syn-volcanic, as many NW–SE-trending dykes in Area V (Fig. 5) do not appear not to have been tilted together with their host lavas. We do not dispute the existence of very narrow-angle unconformities although we did not observe them during our regional mapping of the basalt stratigraphy. There is clear evidence of active faulting (syn-volcanic fault scarps) and subsidence during eruption of the lower part of the Vaigat Formation, and we have also pointed out that there must be an angular unconformity between the Svartenhuk Formation and the Vaigat Formation south of Usuit Kuussuat, proving that there was tectonic activity, including tilting, during volcanism. However, we believe that much of the tilting that affected the Svartenhuk Formation, and therefore also the underlying Vaigat Formation, is the result of later tectonic activity. In the first place, both young dykes and the youngest basalts (those overlying the Arfertuarsuk trachyte) have been tilted locally more than 25°. Secondly, if Geoffroy *et al.*'s interpretation applies to the Svartenhuk Formation, stratigraphic units in this formation should thicken south-westwards as they approach the NW–SE extensional faults, in particular the Arfertuarsuk fault. However, at least as regards the lowest unit in the formation, there is no systematic thickening from north-east to south-west. This unit is 800 m thick on Innerit and north-east of Qooroq where it dips less than 5° towards south-west (see Plate 1, section C–D), 500 m thick in central Svartenhuk Halvø (see Plate 1, section A–B south-west of Usuit Kuussuat), 700 m thick north-west of Tasiusaq, and 850 m thick south-east of Tasiusaq where the dip is 28° towards south-west. Finally, in north-west Nuussuaq there is a 52.5 Ma old comendite tuff in basalts dipping 15° NW (Storey *et al.* 1998), proving that substantial tilt occurred here about 7 Ma *after* the eruption of the main lava succession.

However this may be, we are reluctant to accept Geoffroy *et al.*'s (1997, 1998, 1999) contention that the seaward-dipping basalt areas of central West Greenland provide a convenient model for seaward-dipping reflectors in offshore areas around the North Atlantic. Both western Disko and Svartenhuk Halvø are underlain by continental crust, and the basalts in these areas and for some tens of kilometres to the west offshore are underlain by thick successions of Paleocene and earlier sediments. Inclusions of mudstones occur in the volcanic rocks in west Disko and south-western Svartenhuk Halvø, and offshore sedimentary basins can be extrapolated from both north and south into the area west of Disko and Svartenhuk (Fig. 14). In this setting extension is likely to lead to development of listric faults that level out into detachment zones in the underlying sediments. No such ductile units underlie for example the seaward-dipping reflector sequences off South-East Greenland (Larsen & Jakobsdóttir 1988). Furthermore, dips in the basalts in southern Svartenhuk Halvø and parts of western Disko are steeper than any recorded in offshore seaward-dipping reflectors, and indeed such steep dips can be difficult to image in reflection seismic data without special processing procedures. It would be much more useful to look to Iceland rather than onshore West Greenland to find onshore analogues of oceanic seaward-dipping reflector sequences (Larsen *et al.* 1994).

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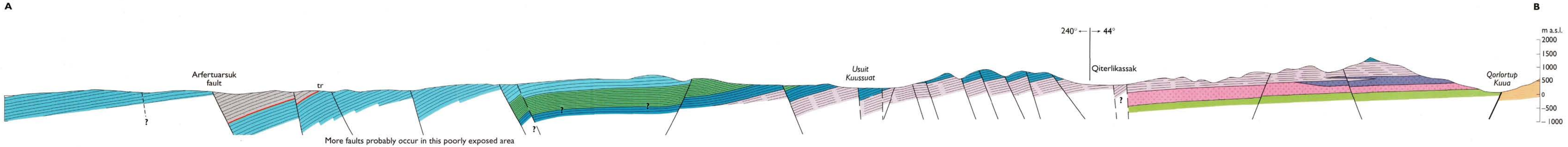
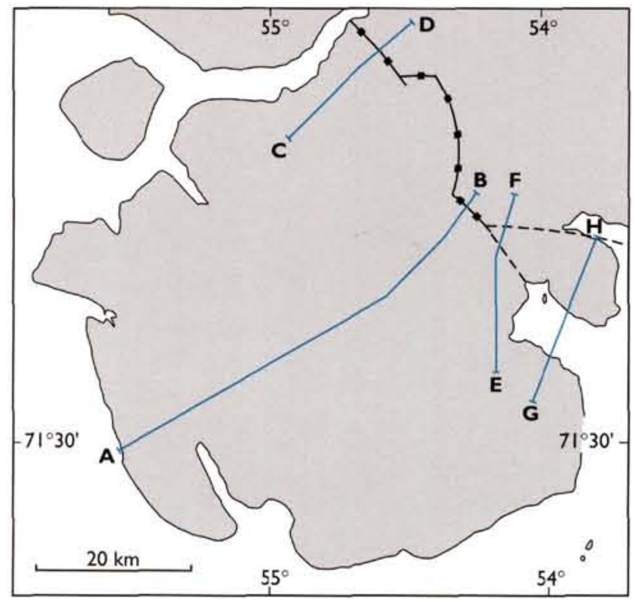
In pocket

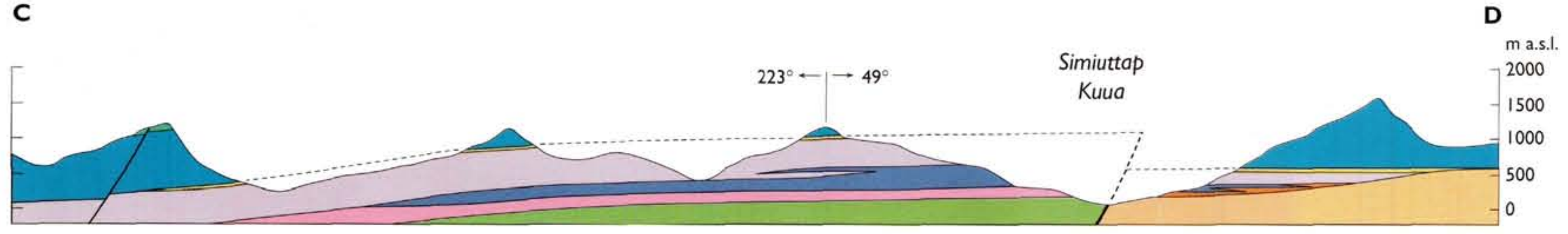
Plate 1. Cross-sections of Svartenhuk Halvø, central West Greenland. Note that the colourless area in section E–F just north of the Cretaceous boundary fault consists of sediments of uncertain age.

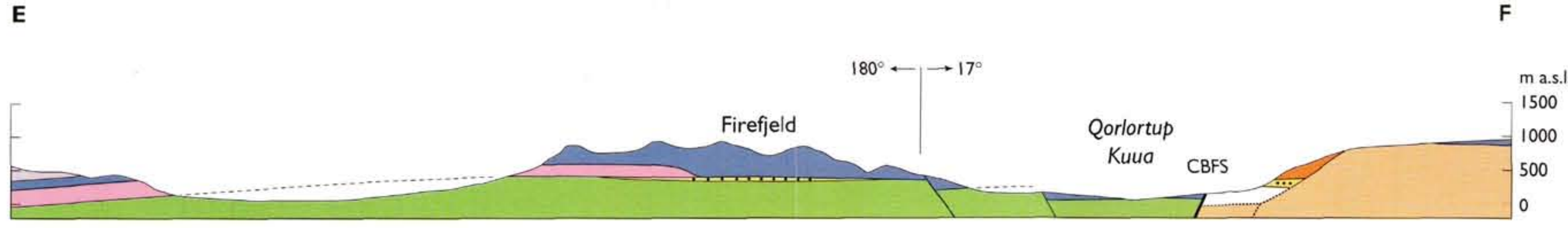
CROSS-SECTIONS OF SVARTENHUK HALVØ, CENTRAL WEST GREENLAND

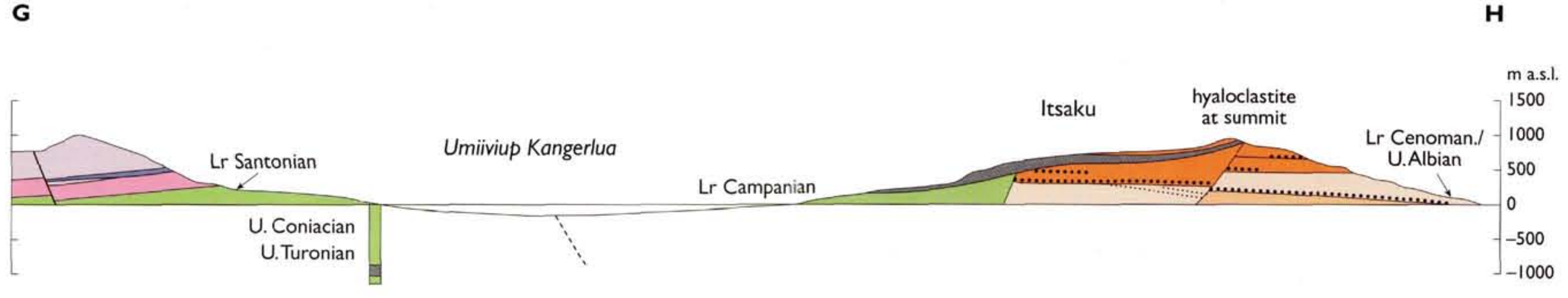
Plate 1











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Volcanic units within the Svartenhuk Formation
tr = Arfertuarsuk trachyte

- 🟨 Intrabasaltic sediments
- 🟪 Tholeiitic picrite lava flows
- 🟦 Tholeiitic picrite hyaloclastite breccias
- 🟥 Kakilisaat Member (sections E–F and G–H); Nerutusoq Member (section C–D); both members present in section A–B

Vaigat Formation

- ∴ Paleocene conglomerate
- 🟧 Upper Campanian/Maastrichtian (?) – Paleocene sediments with conglomerate
- 🟩 Turonian – lower Campanian marine sediments
- ⬜ Upper Albian – Cenomanian deltaic sediments
- ⬛ Dolerite
- 🟧 Precambrian basement

The cross-sections are designed to be read also in conjunction with the Geological map of Greenland, 1:100 000, Svartenhuk, 71 V.1 Nord. Colours used in the cross-sections approximate to those used on the 1:100 000 map sheet with the exception of the upper Campanian/Maastrichtian (?) – Paleocene sediments which are shown in yellowish brown on the cross-sections instead of the green colour used on the map. Note that parallel lines have been drawn within the units shown in Section A–B to bring out structure. They should not be interpreted as indicating unlimited lateral extension of individual lava flows or perfect concordance between all the units.