GRØNLANDS GEOLOGISKE UNDERSØGELSE
RAPPORT Nr. 140

The Geological Survey of Greenland
Report No. 140

REPORT OF ACTIVITIES, 1987

KØBENHAVN 1988

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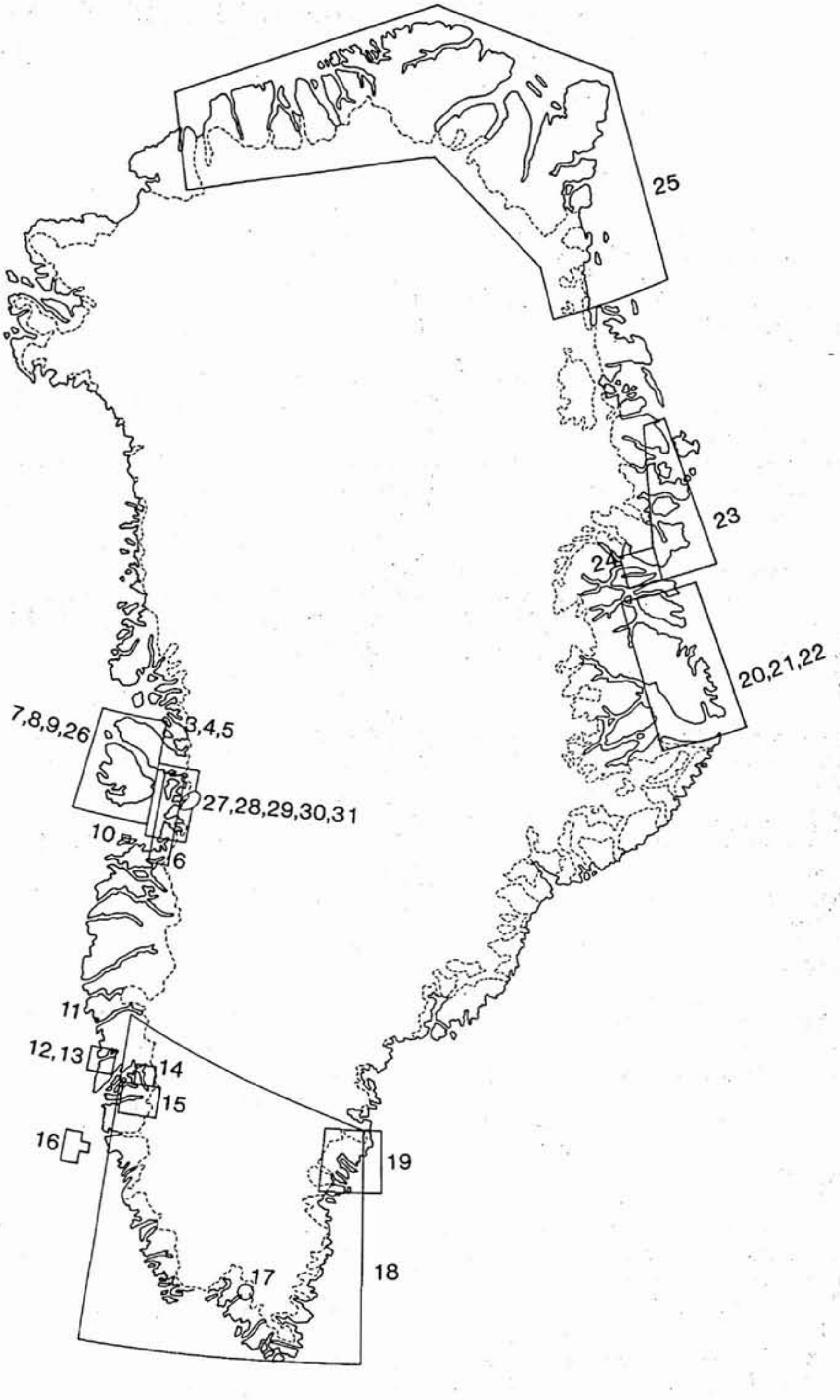

Fig. 1. The Survey's main operational areas in Greenland in 1987. The numbers locate areas discussed in the articles in this report.

CONTENTS

Bold numerals indicate location of areas in fig. 1

GLACIOLOGY

The new Greenlandic orthography will gradually replace the old spelling of geographical names. In this report application of the new orthography is restricted to the names of some of the towns and settlements.

New:	Ammassalik	Nuuk	Narsarsuaq
Old:	Angmagssalik	Nûk	Narssarssuaq
Danish:	-	Godthåb	-

Where no author address is given, the Survey's address in Copenhagen is assumed.

Review of the Survey's activities in 1987

Martin Ghisler
Director

Over the past few years many functions of the Ministry for Greenland have been transferred to the Greenland Home Rule Administration. Following the reorganisation of the government after the general election in September 1987 the Ministry for Greenland was abolished and its remaining functions transferred to other ministries. The Geological Survey of Greenland, together with the Greenland Mineral Resources Administration and the Greenland Environment Research Institute were transferred to the Danish Ministry of Energy. The Minister of Energy, Svend Erik Hovmand, visited GGU in October.

The systematic geological investigation of Greenland and its mineral deposits and petroleum potential continued throughout 1987 – GGU's work comprised geological mapping and reconnaissance surveys combined with more detailed studies in selected areas with emphasis on economic geology.

The scientific and technical staff of 114 continues to be based in Copenhagen. A total of 92 participants were engaged in the field work in Greenland and carried out programmes in both East and West Greenland.

East Greenland

Substantial helicopter-supported field activity continued in two areas in East Greenland involving 40 participants. One group, operating out from a base camp in Hudson Land 200 km to the north of Mesters Vig, studied the geology of the Mesozoic sedimentary basin between Geographical Society Ø and Hochstetter Forland, with special emphasis on an assessment of the hydrocarbon potential. To the south of Mesters Vig a research team financially supported by British Petroleum made detailed stratigraphic studies on selected profiles of the Permo-Triassic sequence in Jameson Land

The other main group in East Greenland undertook regional geological investigations from a base camp near Skjoldungen 300 km south of Ammassalik (Angmagssalik), which was shared with a team from the Danish Geodetic Institute. This group completed the 1:500 000 geological mapping programme of South-East Greenland (map sheet 14), together with a reconnaissance of the area's economic mineral potential.

West Greenland

In West Greenland the systematic geological mapping of the 1:100 000 Fiskefjord sheet (64 V.1 N) was completed. The programme included sampling throughout the area of heavy mineral concentrates from streams, and a detailed study of the scheelite mineralisation at Ivisartoq in the inner part of Godthåbsfjord.

A reconnaissance of the mineral potential of the area north of Jakobshavn, principally in the Precambrian supracrustal rocks was undertaken with special emphasis on sulphides and gold, including the collection of heavy mineral concentrates.

On Disko and Nûgssuaq the investigation of the Tertiary basalts and the associated underlying sediments was continued with emphasis on a regional understanding of the evolution of the area.

Near Jakobshavn, where glaciological investigations have been carried out in connection with the development of hydropower, the topography of the base of the margin of the Inland Ice was mapped with radar-sounding techniques, and the thickness of the ice measured in eight drill holes made by a hot water drill developed in the Survey. The annual mass balance measurements in the area were also carried out.

In addition glaciological and hydrological data were collected from GGU's two field stations near Godthåb and Søndre Strømfjord.

South Greenland

As a continuation of the reconnaissance for noble and base metals in the Precambrian supracrustal sequences in South Greenland, the sulphide mineralisation associated with gold and silver in the Kobberminebugt area was investigated and brought to completion.

The niobium-tantalum deposits associated with syenites around Motzfeldt Sø near Narsarsuaq that were

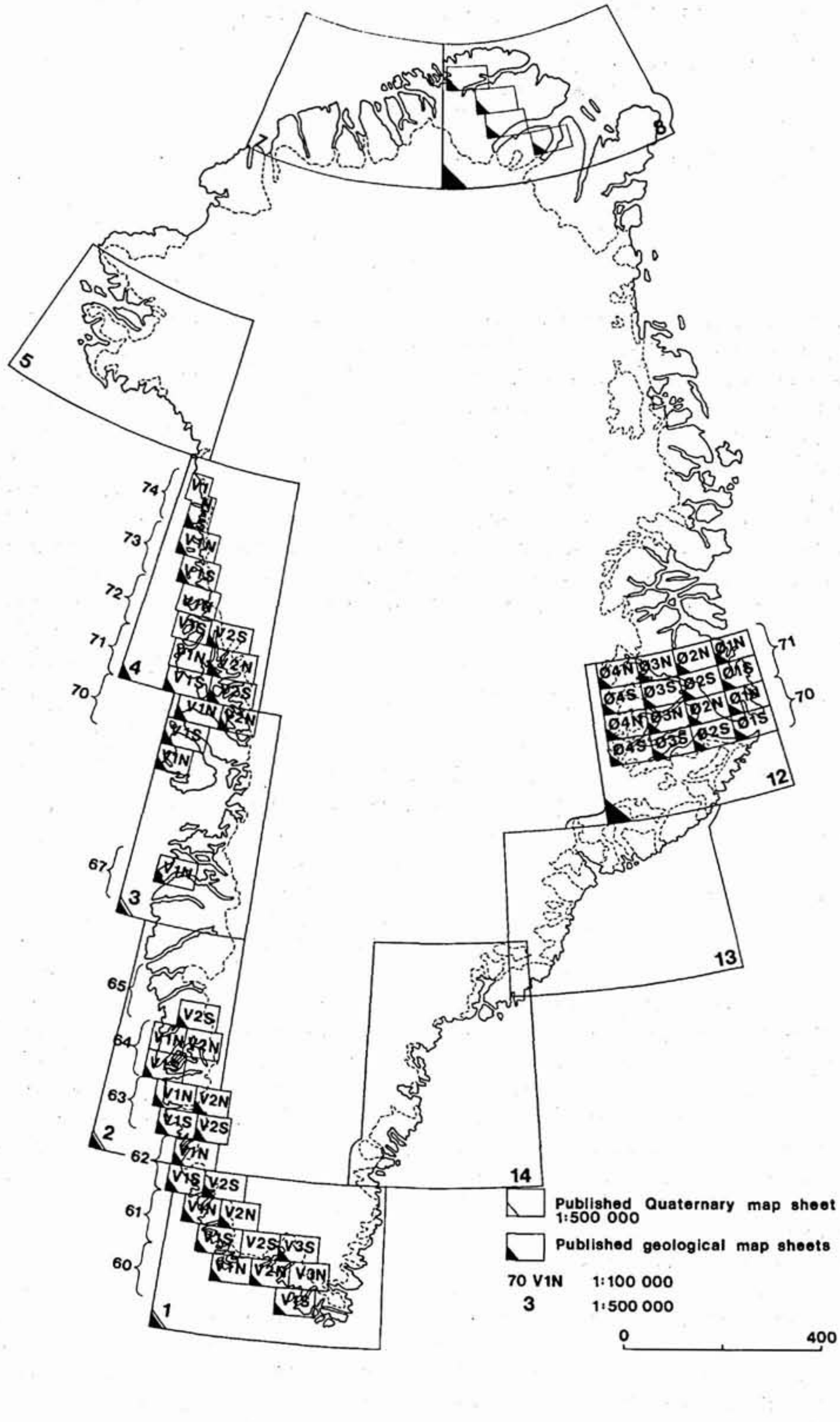

Fig. 2. Map sheets published and in preparation by the Survey (see inside back cover).

identified some years ago were studied in detail. An airborne radiometric survey was followed by a detailed sampling programme over the areas of pyrochlore mineralisation. More than 1000 chip samples were systematically collected from the 1400 m high steep mountain faces by mountaineers. The project was financed by a special grant from the Mineral Resources Administration.

General

GGU inspected the mineral exploration activities of concessionaires at Disko-Nûgssuaq, Ivigtut, Narsaq, Kangerdluarssuk and Nanortalik in West and South Greenland, and at Kangerdlugssuaq and Jameson Land in East Greenland, as well as followed the activities at the Sorte Engel mine at Mârmorilik. The mining activity at Ivigtut was stopped in December 1987. The pit is now filled with sea water and all cryolite from the stock piles has been shipped to Copenhagen. Assistance was also provided to the Mineral Resources Administration in negotiations with applicants for concessions and in the evaluation of concessionaires' reports.

Aeromagnetic measurements, essentially covering the eastern and western part of the Inland Ice south of 66°N, were continued in co-operation with the Geological Survey of Canada and the National Aeronautical Establishment of Canada.

The 'Nordolie' programme funded by the Danish Ministry of Energy since 1984 was completed by the end of the year. The final project report, giving an evaluation of the hydrocarbon potential of central North Greenland, will be published by the Survey.

During the year the 1:500 000 Quaternary map sheet covering South Greenland was printed as well as three geological maps at 1:100 000 from West Greenland (Agpat 70 V.2 N; Mellemfjord 69 V.1 N; Isukasia 65 V.2 S). A special Quaternary map at a scale of 1:125 000 along the coastal area of Jameson Land (East Greenland) was printed as well as a detailed (1:7500) map of the scheelite-bearing rocks at Store Malene near Nuuk. Four Reports, two Bulletins and a Map Sheet Description (Mârmorilik, Nûgâtsiaq, Pangnertôq) were published. As a result of GGU activities 36 contributions appeared in international scientific journals in 1987.

Introduction of new computing facilities at the Geological Survey of Greenland

Leif Thorning

From a cautious start in the use of computers in the early 1970s, the Geological Survey of Greenland has developed complex and varied uses of modern computer facilities for both scientific and administrative tasks. GGU's first computer installation, a noisy TTY connected to the Computing Centre of Copenhagen University by a 110 baud telephone modem, was a self-service facility which was not easy to use. Over the years, first with use of a PDP-10 with just one Tektronix 4014 graphic terminal and later a succession of increasingly powerful PDP-11s with many terminals, GGU's in-house facilities just kept ahead of the ever increasing demand for computer services. At the same time a number of programs for special tasks were developed on external facilities, because they required larger computers or special facilities. In the 1980s the demands on the computer facilities requiring many different types of programs, including word processing, had grown so large that GGU's in-house system could no longer handle them satisfactorily. A major reorganisation was required, and consequently activities were divided between personal computers (PCs; mainly administrative) and a new central computer (mainly scientific). This development took place in late 1986 with the purchase of 17 new personal computers and a new central computer with accessory peripheral equipment. This has allowed an increasing integration of computer methods into GGU's activities. A brief summary is given below.

Hardware

The new central computer facility brought into use in January 1987 is a VAX-8200 system in a Digital network (ethernet with Decservers). Two tape stations and three disk drives are attached to the system. Most terminals are of the VT200 series, but there are also graphic terminals (see below). Several printers, among them two laser printers and a high quality plotter (CalComp 1044GT), provide all necessary types of output. The system includes an interface (PSI, X25) for communi-

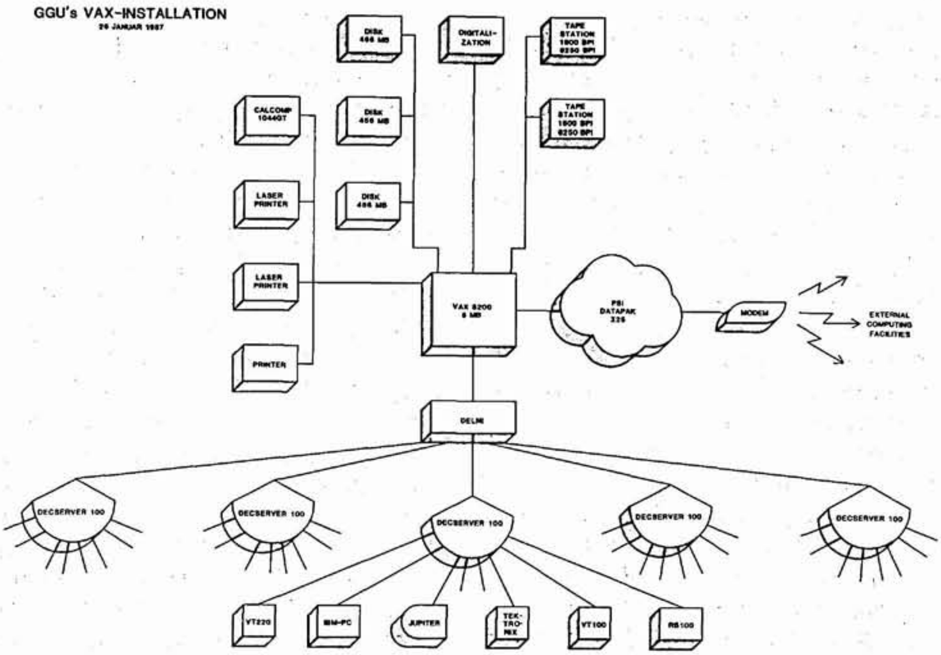

Fig. 1. Schematic diagram of GGU's VAX installation, at 26 January 1987.

cation with other computers. The main components of the system are shown with some specifications in fig. 1. All data and most programs were transferred from the old machine, a PDP-11/44; the change of system caused very little disruption of work. Two main objectives, viz. greater stability and larger capacity, were thus gained almost immediately.

There are several other local computer installations in GGU. These are used for various data-gathering activities in laboratories, e.g. in the chemical analysis laboratory (Hewlett-Packard and PDP-11), the photogeological laboratory (Hewlett-Packard), and the petrophysical laboratory (Hewlett-Packard), and they are all directly or indirectly connected to the central computer to facilitate transfer of data for further processing and interpretation. About 20 personal computers (IBM PC/XT and Rainbow 100+) are connected to the VAX computer and can be used on their own or as VT220 or VT100 compatible terminals. They are used for word processing, minor data bases and other local tasks.

Digitization is performed on a 'stand-alone' microcomputer system with a large Summagraphics table. The PC used for this (Comet 3400) provides a user-friendly interface with local editing and processing of data, and is connected to the VAX so that data can be transferred for further processing and plotting. Other graphical devices are Tektronix 4014 terminals with hard copy units, a Tektronix 4105 with colour printer, and a Jupiter 7+ high-resolution, colour screen soon to be extended with a colour printer.

A considerable amount of processing is still done on external computers, notably on two mainframes in Copenhagen belonging to UNI-C (Danish Computing Center for Research and Education) but also on smaller installations, e.g. IDIMS at the Technical University of Denmark used for image processing. Except for the latter, communication takes place through the public data net (DATAPAK, X25). Most of the geophysical computer work which has the longest history in computing at GGU is in this category. Examples are processing of large amounts of airborne geophysical data at UNI-C using GGU programs, or seismic processing done by contractors on specialized computers abroad.

Software and applications

GGU's software policy operates with three levels of programs, excluding the operating system itself. The highest level includes standard commercial programs from Digital Equipment Corporation (e.g. Fortran and Pascal compilers; Rdb, a relational data base management system; DATATRIEVE, a query and reporting program; FMS, a forms management program; CDD, a common data dictionary system; PSI, a communication program) or from other companies (e.g. GPGS, a subroutine and program library for plotting; BMDP, a statistical package; LEX, a word processing program; SPIDER, a library of image processing routines; NAG, a library of numerical algorithms). The next level is composed of what are termed 'GGU programs', programs originally written by GGU employees or obtained from elsewhere and installed at GGU. These programs are maintained by GGU's computing group, because they are useful to large groups in GGU. Examples are programs for geographical transformations, plotting of maps and geochemical analyses, norm calculations, etc. The third level consists of 'user programs', maintained by the users themselves for their own use. This level includes several programs running on external computers and many sequential file data bases on a personal/departmental basis.

Many and varied jobs, including programs for most types of geoscience interpretation work, are run on GGU's VAX computer by nearly 60 regular users. Three programmers assist the users in different ways and a computer scientist will be added to the staff in the spring of 1988. Both development and production type processing take place.

Trends for the near future

Recently, GGU's needs and obligations for central data bases were analyzed by an internal working group which recommended the construction of central GGU data bases for scientific data. The development of the central data bases will take place in steps, starting with scientific information related to geological samples (sample information, sample description, results of analyses) and later moving on to other types of data, e.g. map-related data with areal extent, glaciological data, etc.

Another main trend which GGU has already embarked upon is the increased integration of map plotting facilities.

An increased integration of the processing of mineralogical/chemical analyses and interpretation, based on results from samples, is foreseen.

Image processing has been used by GGU in various areas. Suitable software will be installed, thus gradually updating GGU's capability for data synthesis, based on information from GGU's data bases and from other sources. In some cases closer ties will be established to GGU's map production facilities. To some extent this has already happened for data types which are well suited for computer handling, e.g. geophysical and geochemical data.

In general, GGU's strategy calls for increasing integration and rationalization of computer application, including transfer of GGU programs from external computer facilities to GGU's VAX-8200. Hardware build-up will take advantage of the network facilities provided by DEC A/S.

Geological reconnaissance in the Precambrian basement of the Atâ area, central West Greenland

C. Knudsen, P. W. U. Appel, B. Hageskov and L. Skjernaa

Geological reconnaissance was carried out in July 1987 as part of regional geological investigations planned for 1986–1992. The field work was carried out from three field camps with logistic support from GGU cutter 'J. F. Johnstrup', a helicopter and a rubber dinghy. The mapping was made on aerial photographs.

General geology

The area has been mapped on a reconnaissance scale (map sheet 1:500 000) and is described by Escher & Burri (1967). Kalsbeek *et al.* (in press) made an isotope-geochemical study of some of the rocks in the area. Steenfelt (1987) carried out a regional stream-sediment geochemical programme, and the mining companies Kryolitselskabet Øresund A/S and Vestgron Mines Ltd. have prospected in the area.

Escher & Burri (1967) divided the rocks in the area into the infracrustal Jakobshavn gneiss and Atâ granite, and the Anap nunâ supracrustals (fig. 1). They described a gradual transition from the mainly granodioritic and locally migmatitic Jakobshavn gneiss to the very little deformed Atâ granite. Escher & Pulvertaft (1976) described the rocks as belonging to the Proterozoic Rinkian mobile belt (a continuation of the Foxe fold belt in Baffin Island) because dome and basin type structures described by Escher & Burri (1967) resemble those of the Rinkian. Escher & Pulvertaft describe the area as separated from the Proterozoic Nagssugtoqidian mobile belt to the south by the Pâkitsoq shear zone.

Rapp. Grønlands geol. Unders. 140, 9–17 (1988)

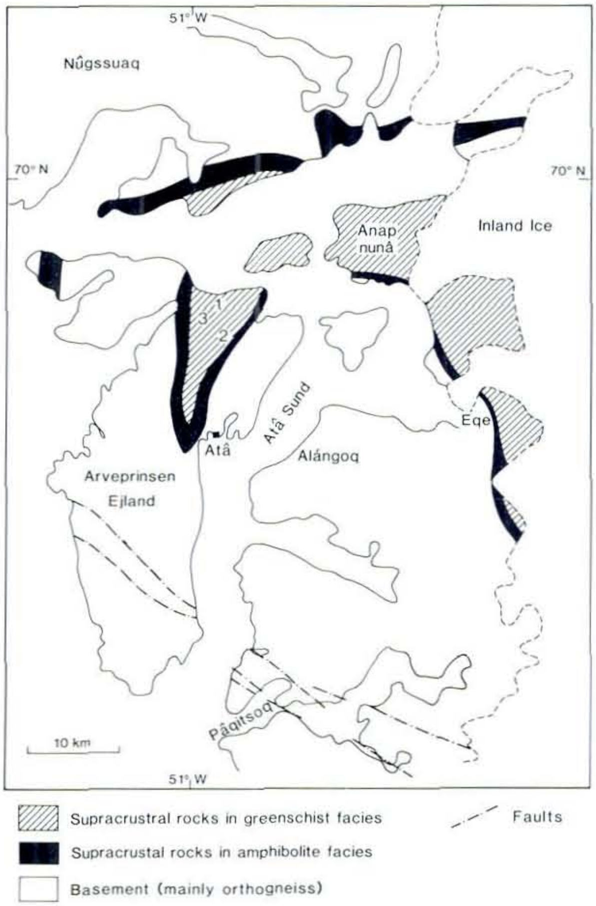

Fig. 1. Generalised geological map of the Atâ area (after Escher & Burri, 1967).

The supracrustal rocks in the area consist of igneous and sedimentary rocks that have undergone greenschist to amphibolite facies metamorphism (fig. 1). A wide range of metamorphosed igneous rock types are represented, ranging from ultramafic rocks to intermediate and acid metavolcanics. The metasediments consist of a variety of siliciclastic rocks, as well as marble, chert and iron-formation. At least two generations of basic intrusive rocks are found in the supracrustals. Abundant massive and disseminated sulphide mineralisations have been found in the supracrustals.

The major part of the fieldwork was in the supracrustal rocks at Eqe and on Arveprinsen Ejland. Further the transition from low strain Atâ granitoids to gneiss and migmatite as well as the fault and shear zones at Pâqitsoq were also studied. However, as only three weeks were spent in the area the following must be regarded as a preliminary report.

The supracrustals

Eqe. Chlorite schist is the most abundant rock type in the supracrustal rocks of this area (figs 4 and 5); it consists mainly of chlorite and plagioclase with varying amounts of carbonate and calcsilicate minerals. Locally the chlorite schist contains carbonate lenses or schlieren, and the schists are cut by quartz and carbonate veins. In the eastern part of the area pillow structures (fig. 2) deformed to varying degrees can be observed, and locally it is possible to determine the stratigraphic way up from the shape of the pillows. Observations indicate that this is towards the west, which means that the supracrustal sequence is inverted. The presence of pillow structures shows that the chlorite schists have been formed at least partly by metamorphism of basic submarine lavas. Locally it is possible to identify relict rims on the pillows, and carbonate or calcsilicate in the space between pillows. These carbonate filled cavities may be a precursor to the carbonate schlieren seen in the chlorite schist.

The chlorite schist is intruded by (meta-)dolerite sills, locally with relic ophitic texture, but where heavily deformed these metadolerites are difficult to differentiate from the metavolcanic mafic schists, and they may be more abundant than indicated on the map (fig. 4). The metadolerites are cut by carbonate and quartz veins and are locally heavily altered.

As indicated in the tentative stratigraphy (fig. 5), metasedimentary horizons are fairly widespread in the supracrustal sequence. In the stratigraphically lower

Fig. 2. Pillow lava. Stratigraphic way up is indicated by arrow.

part there is a thick metasedimentary sequence dominated by quartzites and quartz-feldspar-sericite schists, locally fuchsite-rich or with intercalations of impure, dolomite-dominated carbonates. Several sulphide (mainly pyrite) bearing rust zones occur in these metasediments, often associated with the fuchsite. In the upper part of this sequence a deformed intraformational conglomerate is seen. In the upper part of the supracrustals the metasediments generally consist of intimately interlayered phyllites and quartz-feldspar-sericite schists, locally with graded bedding (fig. 3). These rocks are interpreted as metamorphosed flysch facies rocks. They are interlayered with 10 to 50 m thick sheets of chlorite schist. In the south-western part of the Eqe area, this part of the sequence is characterised by 1 to 5 m thick horizons of iron-formation interlayered with phyllites and quartz-sericite schists.

Marble occurs as horizons up to 30 m thick in the northern part of the area, at the same stratigraphic level as the iron-formation which is, however, not found in this part of the area. In the southern part of the supracrustals carbonate layers only up to 2 m thick are found. In addition to the marble horizons, carbonate is widespread in the supracrustal pile, occurring between pillows as discordant veins and as schlieren in the chlorite schists.

The upper part of the supracrustal sequence is dominated by a homogeneous, grey quartz-feldspar-sericite schist locally with coarse feldspar megacrysts, and interpreted as felsic metavolcanic rocks.

The supracrustals are cut by mafic dykes which are slightly folded.

The westernmost part of the supracrustal area at Eqe is dominated by amphibolites, cut by a light, granitic biotite orthogneiss. This gneiss is assumed to belong to the Atâ granitoids and forms semi-concordant sheets in the supracrustals. These orthogneiss sheets have participated in the isoclinal folding which can be observed in the amphibolite dominated supracrustal rocks. The first occurrence of orthogneiss in the supracrustals can be mapped (fig. 4), but there is a gradual transition from the first occurrence of orthogneiss in the supacrustals to where the orthogneiss constitutes more than 50% of the rock (an arbitrary boundary on fig. 4).

There is a general increase in deformational imprint from east to west, from the almost undeformed pillow lavas near the Inland Ice, through rocks with a well developed schistosity to the highly strained, foliated amphibolite facies rocks adjacent to the orthogneiss in the west. In the central part of the area, the schistosity is refolded in open kink folds often with carbonate veins in the axial plane. In the westernmost part a tightly to

Fig. 3. Metasediments with graded bedding. Stratigraphic way up is indicated with arrow.

isoclinally folded foliation refolded by kink folds can be recognised.

East or north-east dipping 0.5 to 3 m wide breccia zones have been found, mainly in the metasediments west of the iron-formation, but breccia zones are also found in chlorite schists. These zones are sub-concordant, matrix- to clast-supported with phyllite and schist clasts and a carbonate, chlorite or quartz-rich matrix. The breccia contains sulphides (described below).

Arveprinsen Ejland. Chlorite schist is the dominant rock type in the supracrustal belt of this area. The chlorite schist can be divided into metamorphosed basic extrusive rocks and metadiorite/gabbro, and in the major part of the belt these are separate units. In highly strained areas it is however difficult to recognise the relic igneous texture/structure, and hence distinguish these as separate units. The metadiorite/gabbro is generally a medium- to coarse-grained, homogeneous rock consisting of chlorite and plagioclase, often with relic igneous texture. In the north-eastern part of the area, a very light coloured metadiorite/gabbro dominated by coarse plagioclase crystals is common and intrusive into the metasediments in the area. The metadiorite/gabbro is locally cut by carbonate veins, and in the north-

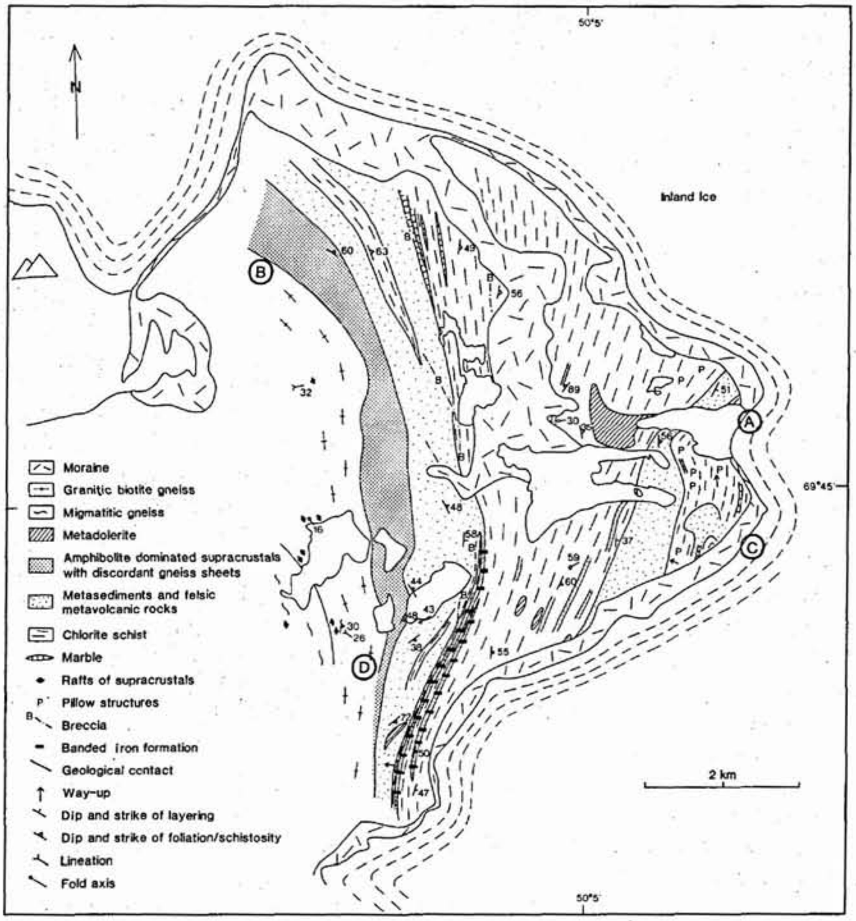

Fig. 4. Geological sketch map of the Eqe area.

western part of the supracrustals it is highly strained and contains carbonate schlieren.

The metasedimentary rocks are generally highly strained, without preserved sedimentary structures, and consist of quartz-feldspar-mica schist to quartzite locally with carbonate layers or carbonate veins. The quartz-feldspar-mica schist is often rusty due to disseminated pyrite or pyrrhotite, and locally small bodies of massive sulphides are found in the metasediments. In the north-eastern part of the supracrustals, almost undeformed rafts of metasediments are found in the metagabbro. Sedimentary structures such as bedding, flame strucures and intraformational clasts/conglomerate can occasionally be observed. In the central and southern part of the supracrustals, felsic rocks containing feldspar phenocrysts are found. These rocks are likely to be metamorphic equivalents to acid and intermediate volcanics.

The metasediments and the metadiorite/gabbro are cut by slightly deformed and metamorphosed basic dykes.

The general trend in the supracrustal rocks at Arveprinsen Ejland is NNE–SSW with a steep to moderate dip towards the east or south-east. In the northern part of the island the trend is ENE–WSW, with overprinting by an east-west trending cleavage. The rocks have participated in large scale, recumbent folding around shallow, north–south trending axes. A crenulation lineation developed on an earlier schistosity is likely to be related to this event. Generally the strain as well as the metamorphic grade increases towards the contacts between the supracrustal rocks and the surrounding orthogneiss. The supracrustals contain a number of highly strained zones within the felsic metasedimentary rocks. In these highly strained zones, the metasediments are very fine grained and have a mylonitic appearance. In the western part of the supracrustals a metasediment horizon locally contains a carbonate rock, and this metasediment horizon also contains a carbonate breccia associated with highly strained rocks. The breccia is locally rich in sulphides (see below). All lithologies including the mylonites are cut by north–south and NNE–SSW trending faults.

The age of the supracrustals. The Atâ granite and the Jakobshavn gneiss have been dated at approximately 2800 Ma, and siltstones from the Anap nunâ supracrustals have a minimum age of approximatly 1800 Ma, and are likly to be of Proterozoic age (Kalsbeek *et al.*, in press). The Atâ granite is, however, intrusive into supracrustals on Arveprinsen Ejland and at Eqe, indicating that these are Archaean. In the Rinkian mobile belt, north of the Atâ area, large thrusts are responsible for interleaving of Archaean and Proterozoic rocks (Pulvertaft, 1986). This may also be the case in the Atâ area, although major thrust zones have not been identified in the supracrustals.

Metamorphism. The major part of the Eqe area is in greenschist facies, but in the western part the supracrustals have been metamorphosed under amphibolite facies conditions. The isograds are discordant to the layering and the amphibolite facies is transgressive in the south. The boundary between greenschist and amphibolite facies is also discordant to the layering in the supracrustals on Arveprinsen Ejland, and there is a general increase in metamorphic grade towards the contact to the orthogneiss. The rocks in the northern part of the area are mainly in greenschist facies whereas the supracrustals at Atâ are mainly in amphibolite facies (fig. 1). This means that carbonate rocks and the carbonate breccia (at 3 on fig. 1) disappear towards the south and are replaced by calcsilicate-rich rocks, common in the western part of the supracrustals near Atâ.

The Atâ granitoids and the gneisses

The relations between the Atâ granitoids and the gneisses and supracrustals of the area have been interpreted in different ways. Escher & Burri (1967) thought that the granitoids (Atâ granite in their terminology) were formed by replacement mainly of the surrounding orthogneisses, but also of the supracrustals.

Kalsbeek *et al.* (in press) suggest that most of the gneisses in the area were formed by progressive deformation of the Atâ granitoids. This idea was supported by their isotopic studies. Age determinations give an age of about 2800 Ma for the undeformed granitoids and about 2670 Ma for the neighbouring gneisses, the latter age being considered to represent the time of deformation. Undeformed dykelets that cross-cut the granitoids yield the same age as the granitoids themselves.

The reconnaissance visit to the area in 1987 revealed that the boundary relations of the Atâ granitoid complex are more complex than formerly believed, and at least some of the relative chronology of the intrusive and deformational events in the area has now been established.

The Atâ granitoids are mainly found in the northern half of Arveprinsen Ejland, at Alángoq and on Igdluluarssuit island south of Anap nunâ. The granitoids include a variety of types ranging from tonalites to granites, and where indications of the age succession are seen, the most granitic types are the youngest. The variation with respect to fabric is from isotropic, through homogeneously foliated and/or lineated to strongly folded. In narrow anastomosing zones, especially along the northeast and east coasts of Arveprinsen Ejland, the granitoids are strongly sheared and have been transposed to mylonites and phyllonites. The isotropic varieties of the granitoids are mainly found in the central parts of the complex, while the more deformed types are developed at the borders of the complex approaching the gneisses and towards the supracrustals. Locally there is a gradational contact between the granitoids and the surrounding gneiss. Intense shearing at the boundary between supracrustals and granitoids is seen.

Relations to the gneisses. A system of up to ten generations of cross-cutting dykelets is an important time marker. The composition of these dykelets ranges from tonalitic to granitic, the grain size ranges from fine grained to pegmatitic and widths from a couple of centimetres to a couple of metres. Although the dykelets obviously have different ages as they cross-cut each other, they seem to share a common place in the overall chronology. Several of the dykelets have characteristic appearances, which make them easily recogniseable from one locality to another, even when they are rather strongly deformed. For example, one generation of fine-grained tonalitic dykelets has rims of white pegmatite. Another generation of granitic to tonalitic dykelets show complex internal structures suggesting a synintrusion shearing parallel to the dykelets. Undeformed dykelets are found to cross-cut:

(1) Migmatitic grey gneisses. The folded migmatitic veins as well as the fabric in the gneiss are cut by the dykelets.

(2) Foliated and lineated Atâ granitoids.

(3) Isotropic Atâ granitoids.

Some parts of the Atâ granitoids either lack dykelets or contain only one or two generations. Either the dykelets intruded only parts of the pluton, or some parts of the pluton were emplaced after the intrusion of (most of) the dykelets. The dykelets have not been seen cutting the supracrustals. Deformed dykelets are found in

Table 1. Relative chronology in the Atâ granitoids

Deformation in parts of the area		Young
Dykelets (up to ten generations)		
Deformation in parts of the area		
Granite	} Atâ granitoids	
Fine-grained tonalite	} Atâ granitoids	
Medium-grained tonalite	} Atâ granitoids	
Deformation		
Veins		
Grey plutonite (now migmatitic gneiss)		
Supracrustals		Old

foliated, lineated and folded gneisses along the middle of the west coast of Arveprinsen Ejland and at the west coast of Alangoq opposite Atâ. In both places there is a transition southwards from isotropic or slightly foliated host rock with undeformed dykelets to gneissic host rock with strongly deformed dykelets. These continue to banded gneisses where the dykelets have been transposed to gneissic bands, and only the most competent pegmatites can be recognised as strongly sheared veins or as rootless folds. These gneisses resemble some of the gneisses in the Pakitsôq area further south. At one locality, small folded pegmatite veins in metasediments are cut by tonalite belonging to the Atâ granitoids.

Summary of chronological relations. Based on the relations mentioned above an incomplete relative chronology is proposed (Table 1). The position of the migmatitic gneiss is somewhat uncertain; it may be considered to belong to an early phase of the Atâ granitoids separated from the others by a deformation. At Eqe a similar migmatitic gneiss contains xenoliths of amphibolite facies supracrustals, but the possibility that parts of the migmatitic gneiss have formed the basement to the supracrustals cannot be excluded. The relative age of possible Proterozoic supracrustals is not indicated in Table 1.

Table 2. Chemical analysis of sulphide-bearing grab samples

Sample number	Au* (ppm)	Ag* (ppm)	Cu† (pct)	Zn* (ppm)	Pb† (ppm)	As* (ppm)	Sb* (ppm)	W* (ppm)	Fe* (pct)	Co* (ppm)
341442	0.10	<5	0.03	<200	<6	4170	2	<8	14.0	1050
341445	3.44	23	1.20	300	<6	359	1	70	6.3	73
341448	0.81	30	1.52	<200	<6	151	1	20	9.3	50
341453	0.05	5	0.15	<200	<6	24	0	120	2.3	26
341489	12.30	44	2.67	630	70	67	4	17	8.7	72
341510	0.62	320	13.10	9900	12200	3060	61	<11	48.0	2890
341511	0.08	44	1.17	55900	3930	7	3	27	54.9	930
341528	0.12	59	0.30	22700	134	291	3	25	39.0	210

Samples from Eqe

341442 Ankerite breccia (matrix supported).
341445 Silicified breccia (with chalcopyrite veins, matrix supported).
341448 - - - - - - -
341453 Brecciated phyllite (clast supported).
341489 - - (clast supported with carbonate and chalcopyrite veins).

Samples from Arveprinsen Ejland

341510 Massive sulphides (pyrite and chalcopyrite with subordinate pyrrhotite, sphalerite and galena).
341511 Massive sulphides (pyrrhotite and sphalerite together with amphibole).
341528 Massive sulphides (layered pyrite and sphalerite together with quartz).

* Neutron activation analysis (Bondar-Clegg, Canada).
† Atomic absorption analysis (average of two analyses).

Shear and fault zones at Pâkitsoq

The Pâkitsoq area south of the Atâ granite complex is of considerable interest as Escher & Burri (1967) suggested that the ESE–WNW trending fault zone seen in Pâkitsoq separates the Egedesminde complex in the south from a northern complex including the Umanak area. This fault zone is postulated to be the continuation of a Canadian fault zone along which a sinistral displacement of several hundred kilometres is suggested (Hoffman, in press). The Pâkitsoq area is a rather monotonous terrain formed by grey biotite orthogneisses of granodioritic to tonalitic composition. Presumably this orthogneiss mass includes elements of the Atâ granite complex. In the southern part of the area an augen gneiss occurs with scattered alkali feldspar megacrysts, up to 5 cm in size. In the country gneisses a little hornblende and garnet may be found. The gneisses have been exposed to migmatisation to varying degrees. The migmatites are cut by deformed irregular pink pegmatites likely to be of the same age as the dykelets in the Atâ granitoids. Subordinate amounts of basic rocks occur in agmatite and amphibolite horizons and in younger sheets of metadolerite.

Three major deformational events are recognised in the the Pakitsoq area:

The oldest. This took place under amphibolite facies conditions and resulted in folding of the migmatite structure and a pre-existing foliation.

Intermediate. This event was a ductile shearing under amphibolite facies conditions. The overall ESE–WNW trending structural grain of the Pâkitsoq area results from this event, and all over the area there is evidence of shearing followed by refolding of older structures. During this event an ESE-trending and moderately plunging stretching lineation was formed and most fold axes are parallel to this direction. The most intense shearing is seen in the interior part of Pâkitsoq, where heavily sheared gneisses, protomylonites and mylonites are possibly related to a north-east dipping thrust complex, with an oblique thrusting from E to ESE. In the mylonites intrafolial older folds refolded by younger folds are seen. The ductile shearing event took place after the intrusion of the metadolerite sheets, which are deformed to subconformable bodies in the gneisses. The metadolerites show boudinage, they have been intensly sheared along their margins, and minor shear zones may be seen in their otherwise well preserved central parts. This event of ductile shear took place presumably at the same time as the youngest deformation seen in the Atâ granitoids.

The most recent. This deformational event resulted in brittle faulting. The faults are subvertical, trend ESE–WNW, and are associated with about 50 m wide zones of cataclastites. The amount of displacement has not been determined but the faults did not introduce any exotic rocks into the area and there are no metamorphic contrasts resulting from the faulting. The faults may be of importance, but at the moment there is no reason to suggest very large scale movements connected to the faulting.

Mineral occurrences in the Atâ area

An iron-formation and three types of sulphide occurrences have been found in the Atâ area. Grab-samples have been analysed by neutron activation and atomic absorption, and the results are listed in Table 2.

Iron-formation. This has been studied in the central and southern part of the Eqe area, where it occurs in three horizons at the lowermost part of a metasedimentary sequence (fig. 5) . The character of the iron-formation changes along strike, from a schistose magnetite rich rock interlayered with phyllites and quartz-sericite schist in the central part (in greenschist facies) to an isoclinally folded, banded garnet-quartz-magnetite rock in the southern part (in amphibolite facies). The iron-formation is up to 5 m thick and can be followed 4 km along strike.

Disseminated sulphides. Supracrustals in all parts of the region contain rust zones caused by disseminated sulphides. These zones are up to 100 m long and a few metres wide, and the sulphide found is mainly pyrite, constituting up to 10 modal % of the following rock types: quartzite locally with fuchsite, quartz-sericite schist, limestone/marble and carbonate rich-graphitic schists, chlorite schist, pillow lava and altered metadolerite.

Massive sulphides. These occur as bodies up to 20 m long and several metres wide associated with felsic supracrustals on Arveprinsen Ejland. The largest occurrence is the 'Anderson showing' (location 1 on fig. 1) which is an approximately 20 m long and 10 m thick lens of massive sulphides occurring together with fine-grained, quartz-rich metasedimentary rocks. Two mineral assemblages are present here, a massive pyrite-chalcopyrite ore with subordinate galena (GGU 341510) and a pyrrhotite-chalcopyrite-sphalerite ore rich in amphibole (GGU 341511). Apart from this, a small lens of massive pyrite with subordinate quartz,

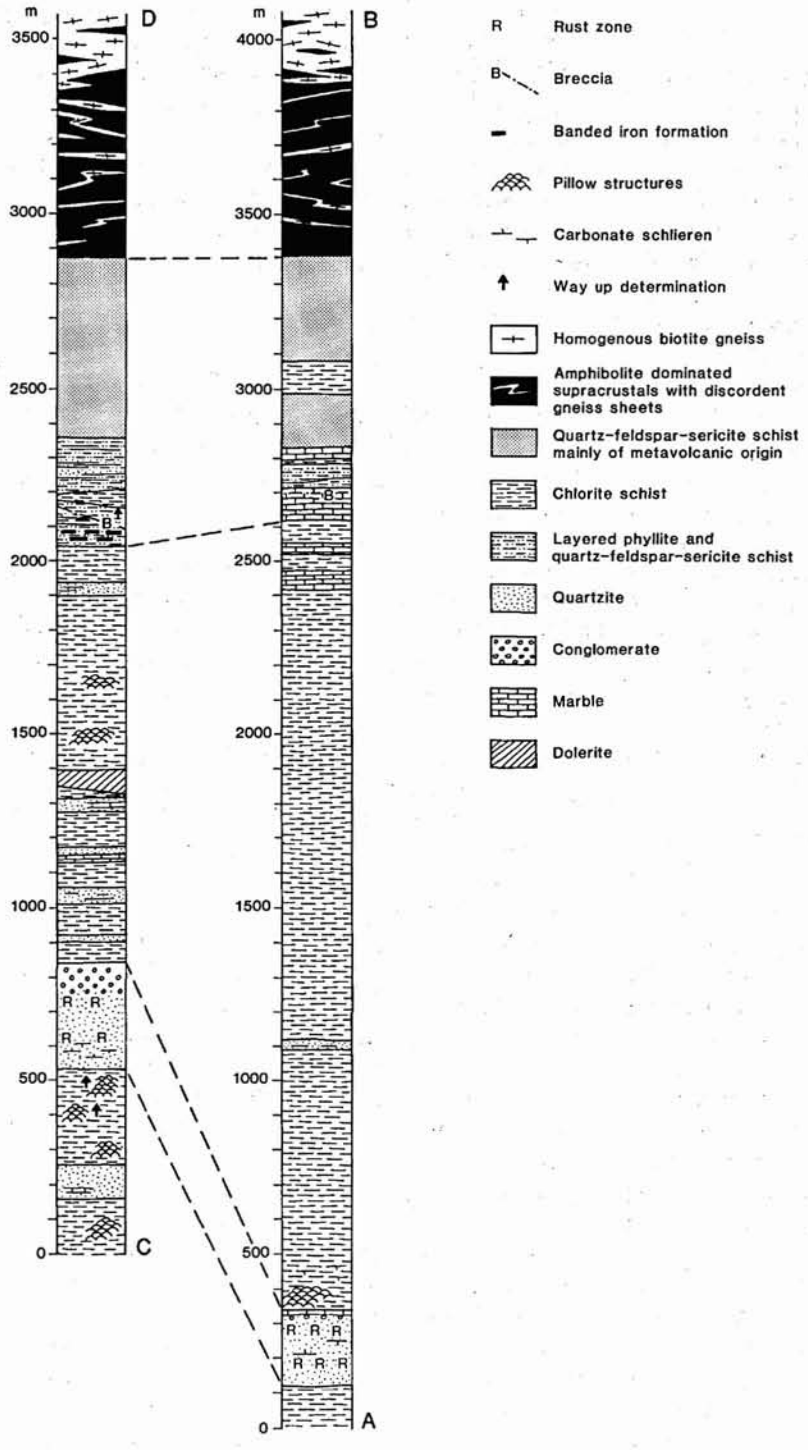

Fig. 5. Tentative columnar sections in the supracrustals at Eqe. Locations of the sections A–B and C–D are indicated on fig. 4. The base of the supracrustal sequence has not been found.

sphalerite rich layers and minor chalcopyrite (GGU 341528) is found at location 2 on fig. 1. This lens is approximately 3 m long and 30 cm wide.

Sulphide impregnations in breccia. Disseminated sulphides are found in breccia zones in the supracrustals. The most interesting of these have been found at Eqe and on Arveprinsen Ejland. The breccia zones are semiconcordant and mainly found in quartz-sericite schist and phyllite, but are also found in chlorite schists and ultramafic rocks (on Anap nunâ). They generally have carbonate (ankerite and calcite), chlorite and quartz in the matrix. The breccia is often slightly porous. The sulphides, generally located in the matrix, consist dominantly of large, anhedral chalcopyrite with minor amounts of euhedral pyrite.

In the Eqe area several chalcopyrite-pyrite bearing breccia zones have been found (fig. 5), and in the southern part of the area a breccia up to 5 m wide could be followed 100 m along strike. The samples GGU 341442, 341445, 341448, 341453 and 341489 (Table 2) were collected in this zone. This breccia is located where Steenfelt (1987) found elevated gold values (up to 43 ppb) in stream sediments.

In the supracrustals at Arveprinsen Ejland a 1 to 5 m wide breccia zone has been followed for about 5 km. The breccia (3 on fig. 1) is locally rich in sulphides, mainly pyrite, but locally massive chalcopyrite and sphalerite occur.

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Progress in geochemical mapping of West Greenland

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Geochemical mapping based on low density sampling and analysis of stream sediment and stream water is part of the mineral resources evaluation programme undertaken by the Geological Survey of Greenland (Steenfelt 1987a, b).

In the field season of 1986 two areas were sampled, the inner Disko Bugt region (1:250 000 map sheet 69 V.2 and part of 70 V.2), and the Godthåb region (map sheets 64 V.1 and 64 V.2).

Sampling and analysis

The sampling was carried out by one field team supported by a Bell 206 helicopter from Greenlandair Charter. At each sample station 200–300 g of stream sediment was collected from four to five sites in the stream bed; a 100 ml stream water sample was taken; and the gamma radiation was measured by averaging five to ten scintillometer readings in the immediate surroundings of the sample site. The sample density is one sample per 20–30 km^2.

The 608 stream-sediment samples were dry-sieved in the laboratory and the < 0.1 mm fraction was analysed by X-ray fluorescence at the Geological Survey of Greenland. A selection of 96 samples were also analysed by Instrumental Neutron Activation at Bondar-Clegg, Canada. The conductivity and the fluoride concentration of the stream water samples were measured in a field laboratory. Table 1 lists the elements and basic statistical parameters for the concentrations determined by X-ray fluorescence and for the water analyses.

The X-ray fluorescence analyses for major and trace elements were made on a Philips PW 1606 multi-channel spectrometer. The analyses were made on glass discs and major elements were determined following the principles of Petersen & Sørensen (1980) while the pro-

Table 1. Statistical parameters for the chemical analyses of the < 0.1 mm fraction of stream-sediment samples from West Greenland

%	min.	max.	mean	s.d.	min.	max.	mean	s.d.
	Disko Bugt 267 samples				*Godthåb 339 samples*			
<0.1 mm fraction of stream sediment								
SiO_2	38.23	73.89	64.57	5.30	38.61	80.85	62.19	4.88
TiO_2	0.22	9.68	0.52	0.59	0.17	2.69	0.52	0.21
Al_2O_3	8.89	24.59	13.63	1.23	7.82	17.67	13.94	0.99
Fe_2O_3	0.70	24.86	4.17	2.09	0.75	14.79	4.88	1.77
MnO	0.04	0.57	0.08	0.05	0.05	0.29	0.10	0.03
MgO	0.47	8.06	1.62	0.97	0.99	25.46	2.52	1.80
CaO	1.72	10.10	3.03	0.71	2.45	9.31	4.36	0.78
K_2O	0.36	2.96	2.00	0.41	0.34	3.01	1.53	0.39
P_2O_5	0.07	2.57	0.19	0.16	0.06	0.48	0.18	0.07
ppm								
V	17	1108	71	71	26	247	87	34
Cr	12	993	144	111	34	921	188	125
Ni	0	484	54	51	2	485	84	67
Cu	0	839	56	74	0	440	45	48
Zn	0	308	55	45	0	581	72	46
Rb	1	170	57	21	0	179	37	25
Sr	122	403	288	53	52	973	295	71
Y	10	76	21	9	0	60	22	7
Zr	64	1478	331	165	17	1905	396	223
Nb	0	77	9	5	0	44	9	4
Ba	44	1150	738	126	85	986	602	117
Stream water samples								
ppb								
F	10	160	16	17.8	10	37	11	3.77
μ ohm^{-1}								
Conductivity	7.8	260.0	28.4	27.7	6.0	92.0	20.6	11.36

X-ray fluorescence analyses by the Geological Survey of Greenland.
min.: minimum value; max.: maximum value; s.d.: standard deviation.

cedure for trace element determinations is under development by Ib Sørensen. The trace element data used here have been corrected for background but not for effects of matrix and line coincidence, and they must be considered preliminary. However, there are no problems with line coincidence for Rb and Cu used in this presentation, and matrix effects are relatively small because of the use of glass discs.

Presentation and preliminary evaluation of results

As a first approach in the evaluation the two regions are looked upon as two (arbitrarily chosen) segments of the Archaean crust. The Godthåb region is established as Archaean (Bridgwater *et al.*, 1976), and the Disko Bugt region is described as Archaean with the exception of some supracrustal sequences of probable Proterozoic age (Kalsbeek, 1986; Kalsbeek *et al.*, in press). The average chemical composition of each segment may be estimated using the values of the stream sediment analyses from Table 1 which lists the range, mean, and standard deviation for the elements with concentrations above the analytical detection limit.

Figure 1 illustrates the chemical composition of the two areas. There appears to be a tendency for the crust at Disko Bugt to be richer in SiO_2, K_2O, Rb, Cu, and Ba, and poorer in Fe_2O_3, MgO, V, Ni, Cr, MnO, Zn, and Zr, relative to the crust at the Godthåb region. The latter is dominated by granulite facies gneisses and is regarded as a deep crustal level. It is therefore sug-

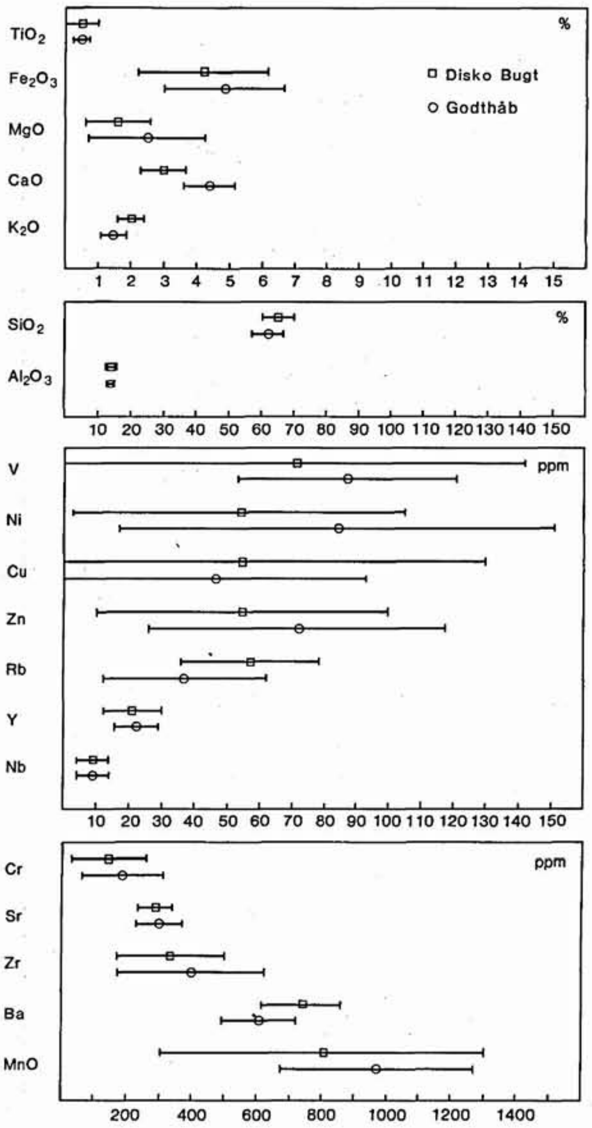


Fig. 1. Graphic presentation of the average chemical composition of the Disko Bugt and Godthåb regions. The bars show the mean standard deviation for the analysis of the < 0.1 mm fraction of stream-sediment samples.

gested that the Disko Bugt region, being richer in 'felsic' components, represents a higher crustal level.

The distribution of element concentrations within the two regions is illustrated in figs 2 and 3. Symbol maps were made for all elements at the scale of 1:500 000. A geochemical atlas containing all the element maps will be published for each region. In the present preliminary evaluation three elements are chosen to illustrate different distribution patterns. Magnesium and rubidium represent elements associated with basaltic and granitic rocks, respectively, and their distribution patterns reflect major lithogeochemical changes. Copper represents an example of an element associated with the sulphide phase which often appears as an indicator of mineralisation in supracrustal rocks.

Previous studies (Steenfelt & Kunzendorf, 1979; and unpublished data) have shown that the chemical composition of the fine fraction of stream sediment in Greenland is close to that of the surrounding bedrock. An illustration of this is presented in Table 2 in which seven samples of the Atâ tonalite (Kalsbeek *et al.*, in press) are compared chemically with five sediment samples from streams draining the Atâ tonalite. Considering the fact that the rock samples were not collected with the aim of finding an average composition of the tonalite, the agreement is satisfactory. A differential weathering effect is noted in that feldspar components like SiO_2, Al_2O_3, Na_2O, and Sr have lower values in the sediment, whereas components typically contained in accessory minerals like P_2O_5, Y, and Zr are enriched in the sediment relative to the rock. In the following, the stream-sediment data are treated as representative of the bedrock, and the distribution patterns displayed by the three elements are compared with major geological units and structural features (figs 2 and 3).

Table 2. Chemical comparison between rock samples and stream-sediment samples derived from the Atâ tonalite

%	7 rock samples	5 stream samples
SiO_2	67.27 – 72.08	59.83 – 67.74
TiO_2	0.22 – 0.47	0.35 – 0.45
Al_2O_3	14.74 – 16.73	12.31 – 13.63
Fe_2O_3	0.25 – 0.81	3.11 – 3.67
FeO	1.24 – 2.30	
MnO	0.03 – 0.05	0.05 – 0.16
MgO	0.45 – 1.06	0.93 – 1.15
CaO	2.28 – 3.41	2.47 – 2.69
Na_2O	4.28 – 5.07	2.95 – 3.56
K_2O	1.19 – 3.23	1.75 – 2.34
P_2O_5	0.08 – 0.15	0.17 – 0.30
ppm		
Rb	43 – 99	43 – 65
Sr	373 – 471	291 – 316
Y	5 – 11	15 – 32
Zr	91 – 163	278 – 381
Ba		574 – 751
V		50 – 61
Cr		85 – 231
Ni		23 – 37
Cu		0 – 110
Zn		23 – 68

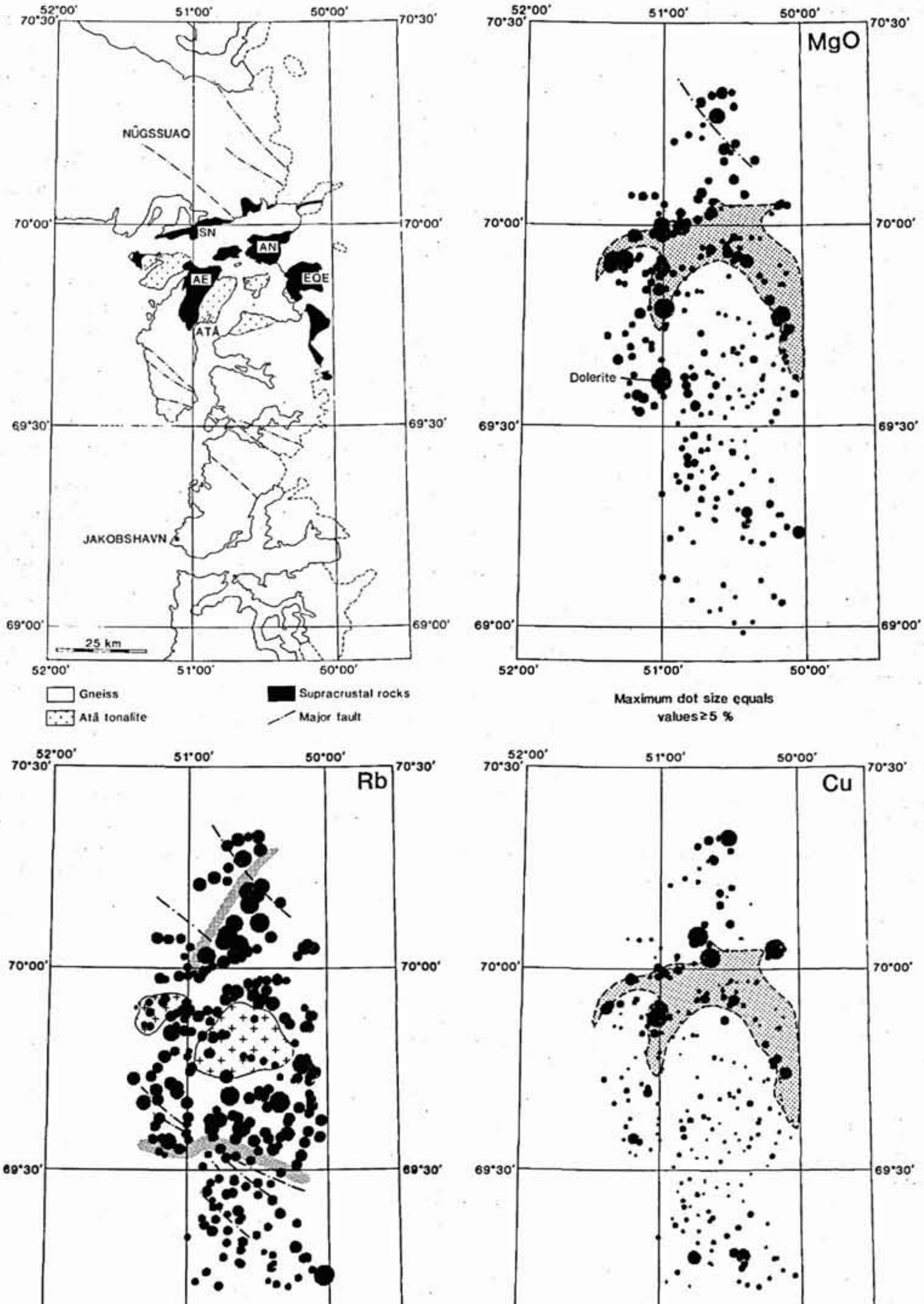


Fig. 2. Simplified geology of the Disko Bugt region and corresponding element distribution maps based on X-ray fluorescence analysis of the < 0.1 mm fraction of stream sediment. The dot size is linear proportional to the element concentration up to a maximum of 5% for MgO, 120 ppm for Rb, and 300 ppm for Cu.

Disko Bugt. Fig. 2 shows that high MgO values delineate the amphibolitic parts of supracrustal sequences in the area. The high value at the east coast of Arveprinsen Ejland marks the presence of a large dolerite dyke. A considerable difference is noticed between the generally low MgO level (< 1.6%) in the gneiss terrain to the south, and the medium to high level (1.5–4.0%) in the gneiss in the north-western corner of Nûgssuaq. The high level may indicate the presence of a dioritic component, or more probable the occurrence of amphibolites included in the gneiss.

The pattern for rubidium shows some distinct changes in concentration level which must be attributed to lithogeochemical changes within the established geological units. The mafic-felsic character of the volcano-sedimentary rocks is reflected in the varying Rb-values in these rocks. Further it is noted that there are three levels in the gneiss areas: a low level (< 50 ppm) in the southern gneiss terrane north and south of Jakobshavn and also in the central part of the Atâ tonalite; a medium to high level (50–90 ppm) in the gneiss surrounding the central Atâ tonalite as well as in the western part of the Nûgssuaq area; and finally a high level (90–120 ppm) in the eastern Nûgssuaq gneiss. Tentative boundaries are drawn between the concentration levels. It is noteworthy that the northern boundary of the southern low-Rb area is not coincident with the major shear zone (A–A in fig. 2) which Escher & Pulvertaft (1976) suggested as representing the boundary between the Nagssugtoqidian and Rinkian mobile belts. The low-Rb boundary follows a set of faults with the same direction, lying *c.* 5 km north of the major zone (cf. Knudsen *et al.*, 1988). The differentiation of the gneiss terrane by the Rb-concentration levels adds a new dimension to the interpretation of the origin and interrelation of the various gneiss units.

The Cu map for the Disko Bugt region shows a low background level and a number of scattered medium and high values mainly, but not exclusively, confined to the supracrustal rocks. The supracrustal rocks are known to contain sulphide mineralisation with chalcopyrite (Knudsen *et al.*, 1988), and this stream-sediment survey confirms that Cu is frequently associated with these rocks. The medium to high Cu values in north-eastern Nûgssuaq coincide with medium to high MgO values and thereby support the assumption of the presence of basic (supracrustal?) rocks.

Godthåb. Fig. 3 shows a belt of high MgO values (3.0–5.0%) with a north-north-east trend following the major mafic sequences of the Malene supracrustal rocks. The north-western corner of the Fiskefjord district is rich in MgO (2–10%) which reflects dioritic to gabbroic rocks and frequent occurrences of ultramafic bodies (Garde *et al.*, 1987; Garde & Marker, 1988).

In the distribution pattern for rubidium a diagonal zoning is observed with a low-level area (< 30 ppm) to the north-west, a central north-east trending zone with high Rb (50–180 ppm) and decreasing values towards the south-east. The low-Rb areas correspond largely to the distribution of granulite facies gneisses, and the high-Rb zone appears as a regional crustal feature transecting gneiss and granite units of different age and origin, i.e. Amîtsoq gneiss, Nûk gneiss and Qôrqut granite (Bridgwater *et al.*, 1976). The high-Rb zone is bounded towards the north-west by major faults and to the south-east the boundary appears to coincide with a major thrust according to recent field work by Nutman & Friend (1988). The high-Rb zone may thus represent a segment of high-level crust in tectonic contact with deeper levels. Data from an airborne radiometric survey support this hypothesis by showing that the Rb-rich zone is also enriched in U and Th (Steenfelt, 1987a).

The fault running across Nordlandet appears to coincide with a change in concentration level for both Rb and MgO which may indicate a significant displacement along the fault. The Cu map shows a low and unsteady

Table 3. Districts with clusters of high values of ore forming elements

Districts	Anomalous metal contents
Disko Bugt	
Oqaitsut	MgO, MnO, V
Arveprinsen Ejland	Au, Cr, Cu, Ni, Zn
Eqe	As, Au, Mo, V, Zn
Anap nunâ	Ag, As, Au, Ba, Ni, Zn
Nûgssuaq south coast	Ag, As, Au, Ba, Cr, Ni, V, W, Zn
Godthåb	
Fiskefjord	Cr, Ni
Bjørneø	Cr, Cu, MnO, Ni, P_2O_5, TiO_2, V
Ivisârtoq	Ag, Au, Co, Cr, Cu, Ni, W, Zn

The anomalies are defined as values above the 98th percentile for both areas. The threshold values used are: Ag 5 ppm, As 80 ppm, Au 5 ppb, Ba 950 ppm, Co 80 ppm, Cr 400 ppm, Cu 170 ppm, MnO 0.8%, Mo 5 ppm, Ni 210 ppm, P_2O_5 0.4%, TiO_2 1.1%, V 185 ppm, W 10 ppm, Zn 170 ppm.

Ba, Cr, Cu, Ni, P_2O_5, TiO_2, V, Zn: X-ray fluorescence analysis by the Geological Survey of Greenland.

Ag, As, Au, Co, Mo, W: Instrumental neutron activation analysis by Bondar-Clegg.

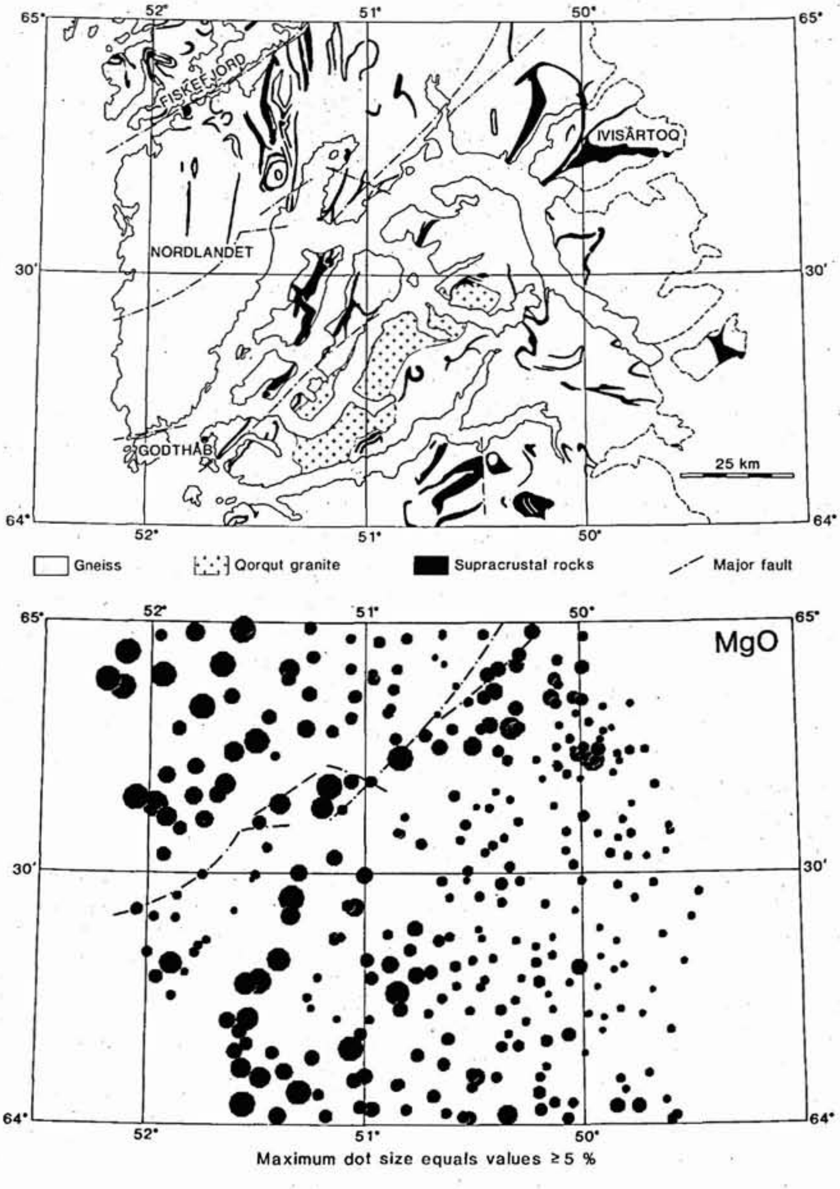


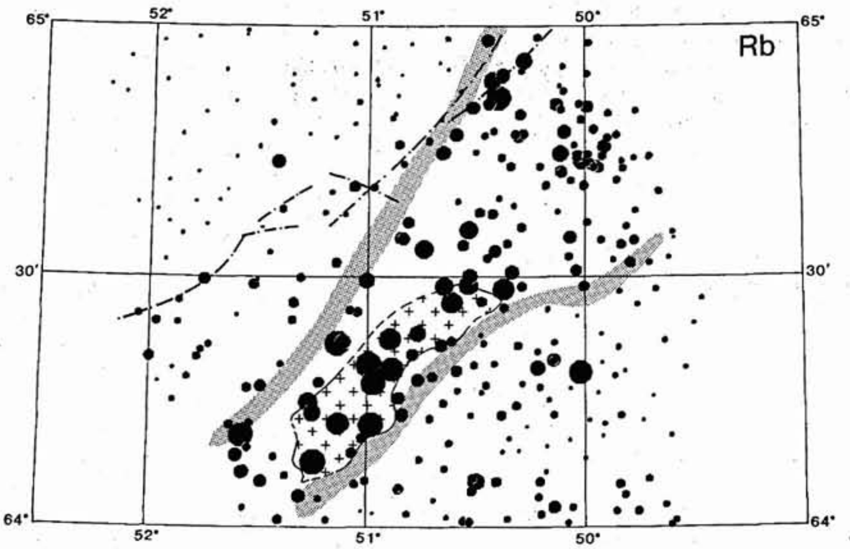
Fig. 3. Simplified geology of the Godthåb region and corresponding element distribution maps based on X-ray fluorescence of

background variation and a few high values. In general there are less high values associated with the supracrustal sequences in the Godthåb region than in the Disko Bugt region.

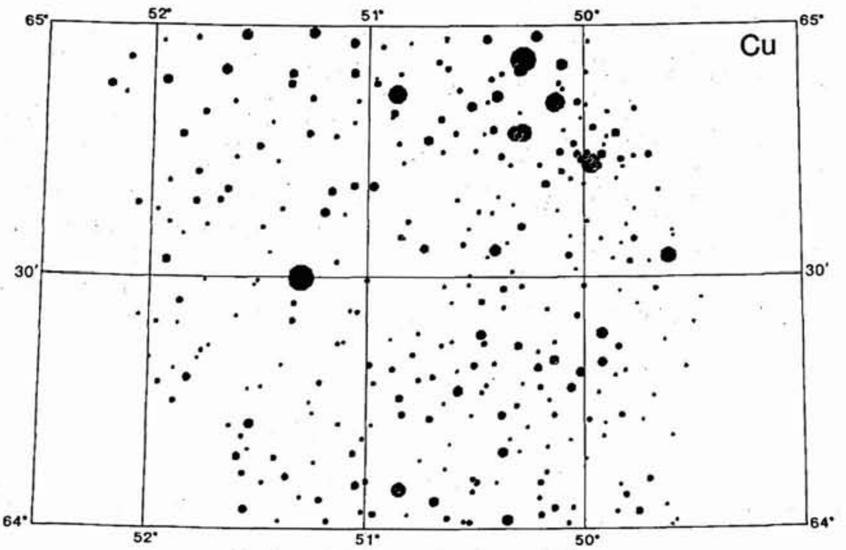
Geochemical anomalies

Samples with unusual chemical composition relative to the geochemical background variation and to the geology of the drainage area may be indicative of mineralisation. The data are not fully evaluated to define and interpret geochemical anomalies but a number of sites containing clusters of high values for one or more ore-forming elements are listed in Table 3 together with the anomalous elements.

In the Disko Bugt region the anomalies are linked to the supracrustal rocks which appear to have a potential for base metals, and Ag and Au. This agrees with Knud-



Maximum dot size equals values ≥ 120 ppm



Maximum dot size equals values ≥ 300 ppm

stream sediment. Symbols as in fig. 2.

sen *et al.* (1988) who report the frequent occurrence of sulphide mineralisation at Eqe and on Arveprinsen Ejland (fig. 2). The results for gold are reported by Steenfelt (1987c). The anomalous sites will be targets for further field work in 1988.

Anomalies in the Godthåb region are scattered and comprise two types. A number of high Cr and Ni values occur in the magnesium rich north-western part of the survey area and are believed to reflect chromite bearing peridotite bodies in the gneiss. Dunite bodies were observed during the sampling by the author and are also described by Garde *et al.* (1987). A few other anomalies (Table 3) are associated with the supracrustal rocks at Bjørneø, and particularly at Ivisârtoq (fig. 3). The mineral potential of the supracrustal rocks of the Godthåb region is treated by Appel (1984, 1986, 1988) based on results from analysis of panned concentrates of stream sediment and detailed geological prospecting. The distributions of gold values are reported by Steenfelt (1987c).

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Stream sediment sampling in the Atâ area, central West Greenland

Peter W. Uitterdijk Appel and Christian Knudsen

In 1982 scheelite was identified in stream sediments in the Nuuk/Godthåb area, about 600 km south of Atâ. Subsequently a regional stream-sediment programme was carried out in the Nuuk area from 1982 to 1987 as a result of which scheelite was found to be quite abundant in the 3800 m.y. old Isukasia supracrustal rocks as well as in the 3300 to 3000 m.y. old Malene supracrustal sequence (Appel, 1988). It was also recognised that there is a close correlation between the number of scheelite grains and the gold content of the heavy mineral concentrates in the Nuuk area (Appel, 1988).

In the Atâ area (fig. 1) extensive outcrops of supracrustal rocks are found. In these supracrustals, which have been metamorphosed to greenschist and amphibolite facies, abundant sulphide-rich horizons are found, as well as sulphide-bearing breccia zones with appreciable gold contents (Knudsen *et al.*, 1988).

During the 1987 field season geological reconnaissance mapping was carried out in two of the supracrustal areas and the gneiss-granite complex enclosing the supracrustal rocks was investigated (Knudsen *et al.*, 1988). A limited programme of stream-sediment sam-

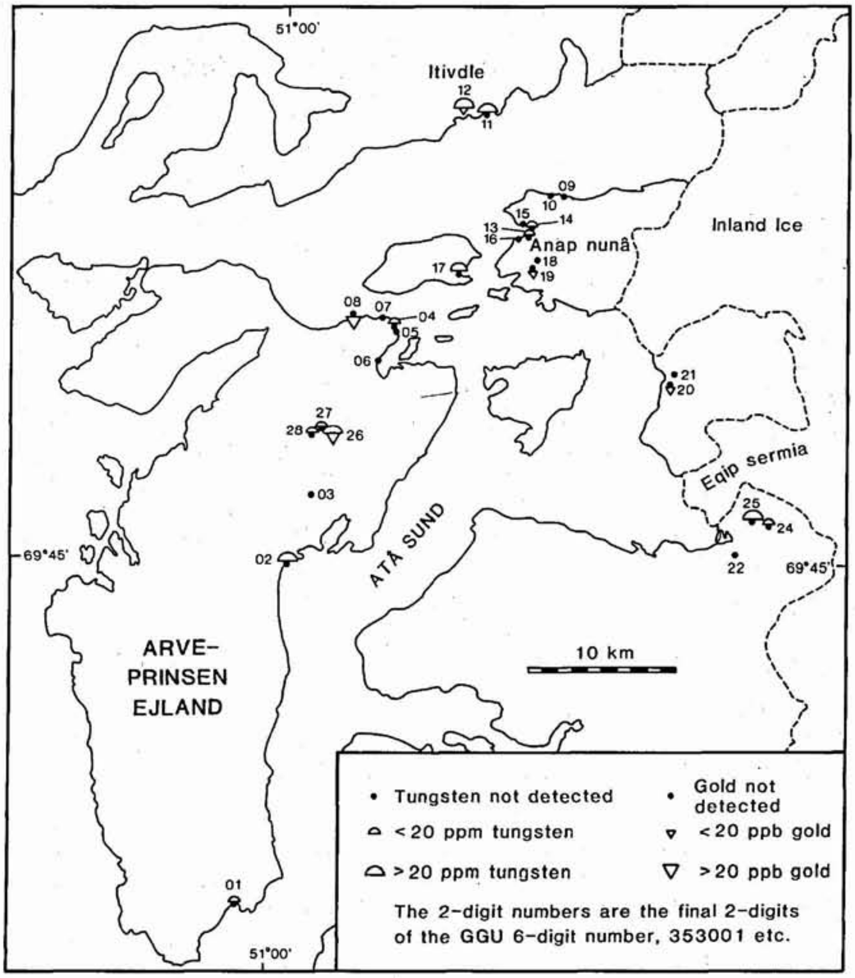

Fig. 1. Map of the Atâ area showing the sample sites with tungsten and gold contents indicated.

pling was carried out in the Atâ area, where stream-sediment samples were collected in all the main streams draining the supracrustal rocks.

Sampling programme

The stream sediments collected in the Atâ area consisted of coarse gravel and sand collected in sieves which contain about 3 kg of sample material. The material was passed through a sieve with 1 mm holes. The fines were measured by volume and then concentrated by panning.

The heavy mineral concentrates were dried and the number of scheelite grains counted under ultra-violet light (except samples 353024 and 353025). The results are listed in Table 1.

Results

The heavy mineral concentrates were analysed by neutron activation by Bondar-Clegg, Ontario, Canada for gold and 33 other elements. The tungsten, gold and barium contents are listed in Table 1. The analytical results for the following elements are available on request: Na, Sc, Cr, Fe, Co, Ni, Zn, As, Se, Br, Rb, Zr, Mo, Ag, Cd, Sn, Sb, Te, Cs, La, Ce, Sm, Eu, Tb, Yb, Lu, Hf, Ta, Ir, Th and U.

There is fairly good agreement between the observed number of scheelite grains and the tungsten content (Table 1), indicating that scheelite is the common tungsten-bearing mineral. There are, however, three notable exceptions, samples 3530001, 353008 and 353017. In samples 353001 and 353017 the tungsten mineral could be wolframite, whereas we cannot explain the discre-

Table 1. Analyses of heavy mineral concentrates from stream sediments in the Atâ area

	Scheelite/l	W ppm	Au ppb	Ba ppm	Sieves	Vol
353001*	0	16	<17	450	3	0.2
353002	10	65	<5	320	3	0.5
353003*	6	<7	<5	270	3	0.6
353004	3	15	<5	250	3	0.3
353005	0	<6	<5	370	3	0.25
353006	2	<5	<5	420	3	0.45
353007	3	<2	<5	460	3	0.3
353008	21	<9	37	280	3	0.35
353009	0	<7	<5	<100	3	0.56
353010	0	<6	<5	650	3	0.5
353011*	30	120	<5	4300	3	0.51
353012*	64	81	16	240	3	0.26
353013	1	7	<5	260	60	5.21
353014	0	13	<13	<100	4	0.2
353015	0	<5	<5	500	4	0.42
353016	0	<5	<5	430	4	0.09
353017	0	24	<17	<100	4	0.35
353018	0	<4	<5	730	3	0.1
353019	0	<6	19	320	3	0.2
353020	0	<8	15	1400	6	0.5
353021	1	<5	<5	380	20	1.45
353022	0	<6	<5	<100	2	1.51
353024		<9	<5	440	3	0.2
353025*		44	<5	260	3	0.2
353026*	32	93	32	290	6	0.6
353027*	6	10	<5	330	6	0.55
353028*	0	7	<5	440	3	0.35

* Samples collected in amphibolite facies.

pancy between 21 grains of scheelite and less than 9 ppm tungsten in sample 353008.

The gold contents in the heavy mineral concentrates are mostly too small to be detected. However, a few samples do contain interesting amounts of gold, and there is apparently some correlation between the gold content and the number of scheelite grains in the heavy mineral concentrates.

It should be noted that some stream sediments were collected in streams draining amphibolite facies metamorphosed supracrustal rocks; 353001, 353003; 353011–353012; 353025–353028.

Conclusion

This limited stream-sediment programme indicates that economically interesting tungsten occurrences may be found in the supracrustals of the Atâ area. Most of the scheelite apparently occurs in the supracrustal rocks which have undergone amphibolite facies metamorphism.

From this work and from the abundance of scheelite previously discovered in the supracrustal rocks further south (Appel, 1988) it is concluded that scheelite is probably more common in Precambrian supracrustal sequences than hitherto realised.

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Reconnaissance aeromagnetic survey east of Disko Bugt, central West Greenland

Leif Thorning

The Geological Survey of Greenland plans a major aeromagnetic survey in the Disko Bugt region over the coming years. It has already been started with some introductory work in 1986 and 1987 (Knudsen *et al.*, 1988). In this context, closed-file geophysical data were reviewed at GGU, and it soon became clear that good quality regional aeromagnetic data were lacking from the area and that the geological investigations would benefit from such data. Funds were not available for a systematic survey over the entire area, but in April 1987 a situation arose in which at least part of the area could be surveyed.

Due to the change of plans for the aeromagnetic surveying of the GICAS project, described by Thorning *et al.* (1988), it was possible to include one survey flight east of Disko Bugt. The measurements were made from the National Aeronautical Establishment aircraft (C-FNRC), which is a well equipped Convair-580 with very good navigational capabilities and a 3-axis magnetic gradiometer.

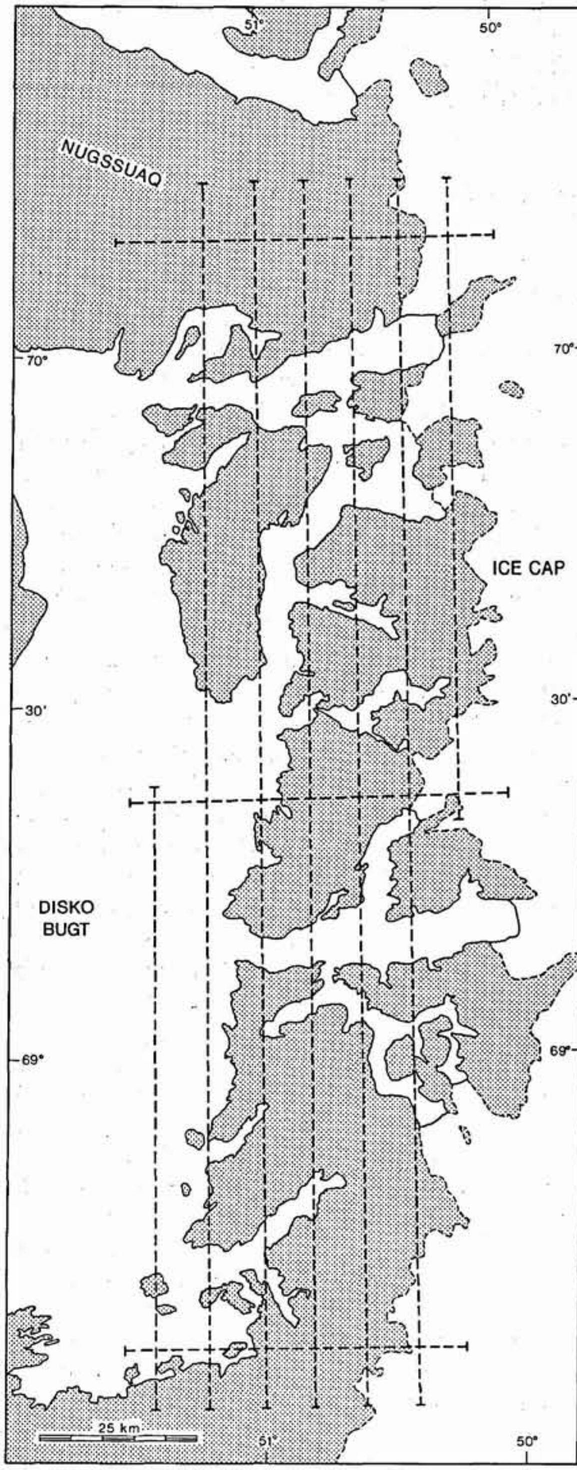

Fig. 1. Aeromagnetic lines flown in April 1987 in central West Greenland. Altogether approximately 1600 line km of data (total field and three gradients) were acquired.

The survey shown in fig. 1 was obtained in one 5 hr flight out of Søndre Strømfjord and comprises seven north–south lines, spaced 10 km apart and three east–west tie-lines approximately 80 km apart. The flight altitude was 1000 ft above ground, sometimes levelling out at 2000 ft over the narrower fjords and valleys. Approximately 1600 line km of total field data and gradients in three directions were obtained. The flight path was recorded on video for later navigation checks.

It was possible to examine the analogue recordings in the field, and these showed many well-defined anomalies, many of which could be followed from line to line. The digital data are still being processed at the National Aeronautical Establishment, before delivery to GGU where they will be compiled. It is hoped that, although the line spacing may be too wide for a faithful rendering of all the smaller scale anomalies which will then have to be interpreted on a profile basis, it will be possible to show at least the larger scale anomalies, corresponding to regional trends, on a magnetic anomaly map. In 1988 the sources of these anomalies will be checked on the ground.

Acknowledgements. The flexibility of my co-workers in the GICAS project made it possible to include the Disko Bugt survey in the tight schedule for the 1987 northern arctic operation. The costs of the operation were defrayed by GGU.

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Investigations of Tertiary volcanic rocks along the south coast of Nûgssuaq and in eastern Disko, 1987

Lotte Melchior Larsen and Asger Ken Pedersen

As a continuation of an integrated study of sedimentary and volcanic facies in the Cretaceous to Tertiary West Greenland basin (G. K. Pedersen, 1987; A. K. Pedersen & Larsen, 1987) early Tertiary volcanic rocks were studied in 1987 along a NW–SE trending composite section, about 120 km in length, on Nûgssuaq and Disko. The study attempts to establish and describe lithostratigraphic volcanic units in the Tertiary volcanic formations, and through a combination of field mapping, photogrammetry and geochemistry to establish chronostratigraphic horizons through the early Tertiary deposits of the region. In this respect it is essential to identify the same eruptive units as subaerial lava facies and as subaqueous lava or hyaloclastite facies, and to trace subaerial tuffs throughout the area.

In the first part of the season localities along the Vaigat coast of Nûgssuaq from Kugssinerssuaq in the east to Nûssap qáqarssua in the west were investigated. In the second part of the season very poorly known areas in the western and southern part of the Kvandalen region on east Disko were investigated. The field work was supported by the Arctic Station in Godhavn and its cutter *Porsild* as well as by GGU's cutter *J. F. Johnstrup*.

South coast of Nûgssuaq

The localities investigated are situated along the western part of the south coast of Nûgssuaq (fig. 1). Information on the volcanic rocks along this part of the section is given by Steenstrup (1900), Koch (1959), Henderson (1973), Münther (1973), Clarke & Pedersen (1976) and Henderson *et al.* (1976), and the section is covered by the geological map sheet 1:100 000 Qutdligssat. In the investigated area the volcanic rocks are very well exposed in up to 1700 m high mountain sides, but large parts of these are inaccessible due to the steep topography. The basal parts of the hyaloclastite deposits overlying sediments are, however, usually accessible. The volcanic rocks of the section belong to the Vaigat Formation except for some high mountain tops. A detailed lithostratigraphy for the Vaigat Formation has been established for the nearby parts of northern Disko (Pedersen, 1985), and attempts were made in 1987 to extend this stratigraphy to Nûgssuaq.

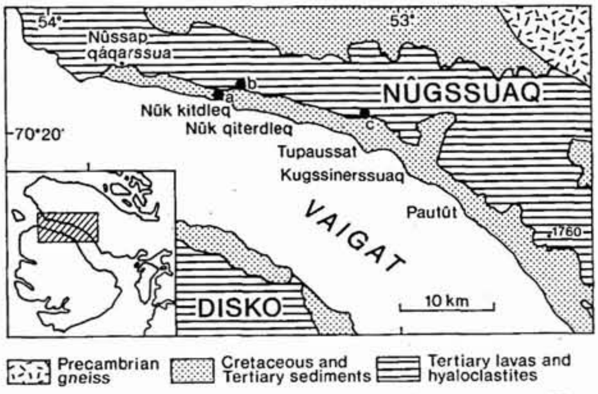

Fig. 1. Investigated area on the south coast of Nûgssuaq. The positions of the three profiles shown in fig. 2 are indicated by a, b and c.

The Vaigat Formation on Disko is divided into six members which were formed by two major igneous events. The first event gave rise to the voluminous picritic volcanics of the Naujánguit Member and two minor sediment contaminated volcanic systems therein, the Asuk and Kûgánguaq Members, while the Qordlortorssuaq Member basalts constitute the waning phase of the event. Local erosion and the formation of a widespread water-filled basin (subsidence) took place between the two igneous events. The second event gave rise to the voluminous picritic volcanics of the Ordlingassoq Member within which are enclosed the alkaline lavas of the Manîtdlat Member.

During the field work on Nûgssuaq special attention was paid to locating the boundary between the rocks from the two major igneous events recognized on Disko, i.e. the base of the Ordlingassoq Member.

The following describes the lower hyaloclastite deposits and subaerial lava flows from the first igneous event, followed by the upper hyaloclastite deposits and subaerial lava flows from the second igneous event.

Lower hyaloclastite deposits. Thick hyaloclastite deposits laid down in a water-filled basin form the basal part of the volcanic sequence from Nûssap qáqarssua to Tupaussat. The deposits in the basin consist in upward sequence of mudstone, hyaloclastite tuffs and foreset-bedded hyaloclastites, the latter being the dominant lithology. The hyaloclastite sequence just west of Nûk kitdleq (fig. 2a) is the thickest of its kind (*c.* 650 m) on Disko and Nûgssuaq and exemplifies the development in the basin.

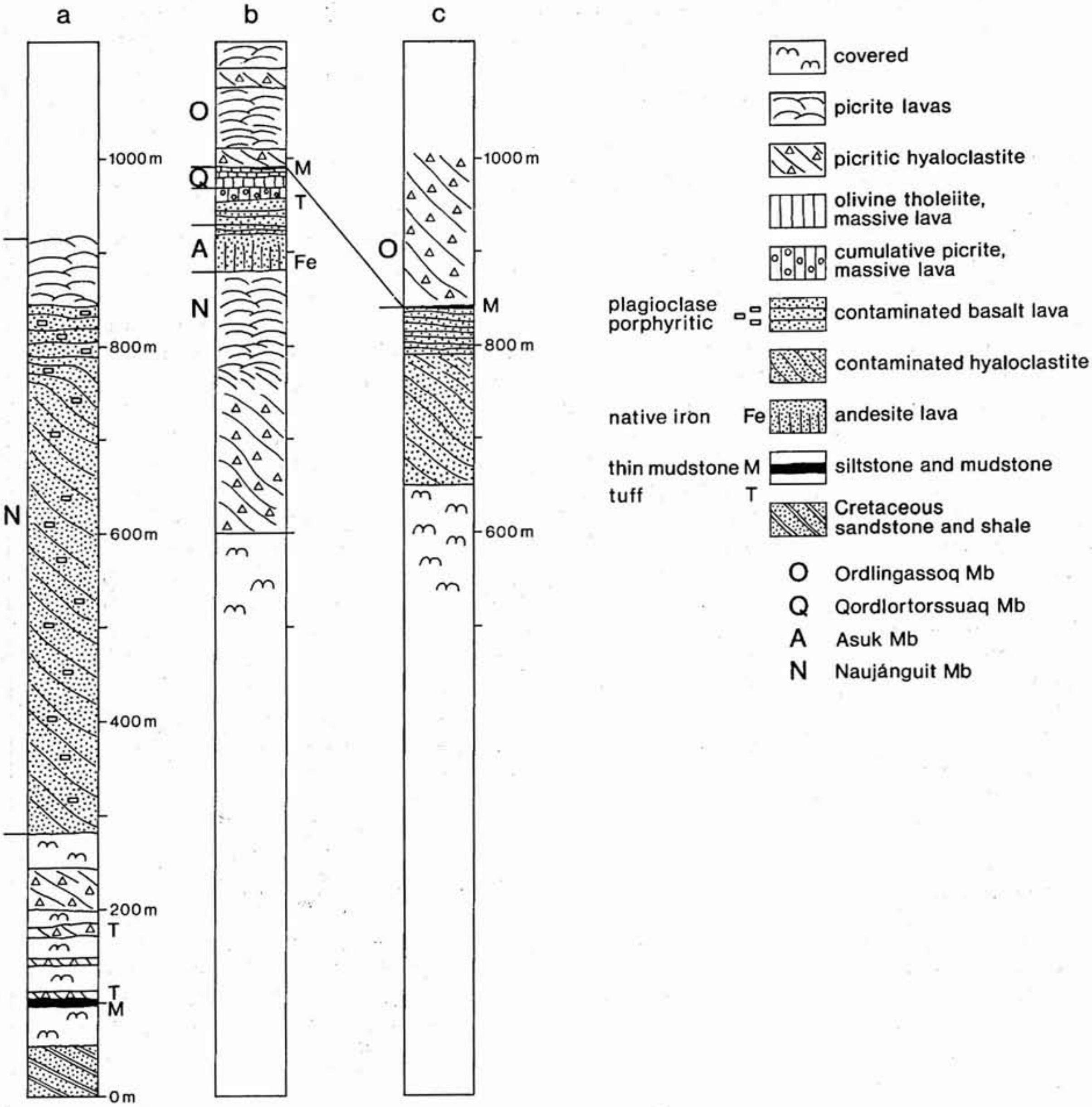

Fig. 2. Profiles through the Vaigat Formation on the south coast of Nûgssuaq. The location of the profiles is shown in fig. 1.

The lower part of the sequence is *c.* 140 m thick and is seen in slightly dislocated or isolated exposures. At the bottom *c.* 5 m mudstone is exposed which is finely intercalated with hyaloclastite in the upper metre. This is covered by a sequence of sandy to silty hyaloclastite tuff units with fining-upwards structure, intercalated with coarser units with pillow fragments. The hyaloclastites originated from picritic magma of a type characteristic of the Naujânguit Member on Disko.

The upper part of the sequence is exposed in near-vertical cliffs. It is about 500 m thick and is composed of regularly foreset-bedded coarse hyaloclastites with pillow fragments of olivine + plagioclase porphyritic contaminated basalt. These hyaloclastites originated from a sequence of yellowish-brown subaerial lava flows which forms a marker horizon within the predominantly grey picritic lava flows towards the west.

The hyaloclastite tuffs in the lower part of the sequence are interpreted as deposited from suspension flows possibly from volcanic sources many kilometres away. In contrast, the foreset-bedded hyaloclastites were formed locally, where subaerial lava flows entered the sea. The deposits gradually filled in the basin, prograding from the west and north-west.

Contaminated volcanic rocks with native iron. In the steep mountain side north of Nûk kitdleq a sequence of sediment-contaminated volcanic rocks is exposed (fig. 2b). A striking feature is an up to 45 m thick rusty brown weathering lava flow with native iron which rests on a sequence of picritic lavas. The brown lava is an erosion resistant columnar jointed andesite with disseminated graphite and native iron. It was first observed on the coast and identified through the study of scree blocks by Steenstrup (1883). The lava is exposed over a distance of 1.2 km and thins rapidly, and stops towards the east. Heaps of strongly graphitic scoria occur at the base of the flow and are rich in xenoliths of magma-modified shale and of a non-graphitic vesicular basalt. A sequence of grey-brown lava flows of contaminated basalts cover the andesite lava or are dammed up against it. Due to lithological similarities these volcanic rocks are tentatively correlated with the Asuk Member on Disko.

Table 1. Chemical compositions of contaminated and uncontaminated volcanic rocks from the Vaigat and Maligât Formations

	1	2	3	4
SiO_2	50.02	45.13	48.73	49.18
TiO_2	0.98	1.28	2.51	1.42
Al_2O_3	12.59	11.49	14.13	14.56
Fe_2O_3	1.12	2.56	4.23	2.55
FeO	8.66	9.02	8.18	8.00
MnO	0.17	0.19	0.20	0.17
MgO	14.94	17.84	7.15	9.66
CaO	8.03	10.54	11.19	10.42
Na_2O	1.38	1.25	2.29	1.65
K_2O	0.29	0.06	0.31	0.24
P_2O_5	0.13	0.13	0.24	0.15
volat.	1.73	0.61	0.60	1.63
	100.04	100.10	99.76	99.63

1. GGU 340711, Vaigat Formation: Contaminated Mg-rich basalt, pillow fragment from the pillow breccia formed late in the first volcanic event. Tupaussat, south Nûgssuaq.
2. GGU 340755, Vaigat Formation: Picrite, pillow fragment in pillow breccia from the Ordlingassoq Member (second volcanic event). Tupaussat, south Nûgssuaq.
3. GGU 340844, Maligât Formation: Olivine + plagioclase porphyritic basalt, pillow fragment in pillow breccia from the Rinks Dal Member. Westernmost Kvandalen, Disko.
4. GGU 318829, Maligât Formation: Contaminated Mg-rich basalt lava from the Niaqussat Member. Peak 1240 north of Kvandalen, Disko.

Olivine porphyritic basalt lavas. A marker horizon above the inferred Asuk Member at Nûk kitdleq is formed by 30–40 m of prominent erosion-resistant grey to dark grey lava flows (fig. 2b). Towards the west this horizon is situated in a subaerial lava sequence, while towards the east the horizon underlies the eastward thickening upper hyaloclastite deposits, until the lavas in the horizon either disappear or develop into hyaloclastite facies 6 to 8 km further east. The lavas consist of olivine-poor tholeiitic basalt and olivine cumulative tholeiitic basalt. These lava flows formed in a period of volcanic stagnation, and the horizon is correlated with the Qordlortorssuaq Member which completed the first volcanic event on Disko.

Contaminated olivine porphyritic basalts at Tupaussat. A sequence of light yellow-brownish weathering pahoehoe lavas of contaminated basalts forms subaerial lavas just below the prominent upper hyaloclastite deposits, stretching from just west of Nûk qiterdleq to Tupaussat. The sequence is coeval with or slightly older than the lava flows assigned to the Qordlortorssuaq Member further west. In the corrie north-east of Tupaussat (fig. 2c) the sequence consists of more than 100 m of hyaloclastite rich in decimetre-sized pillows, covered by 40 m of pahoehoe lavas, while a few kilometres further to the east the whole sequence is in hyaloclastite facies. The rocks are olivine porphyritic contaminated basalts (Table 1, no 1), derived from picrite magma.

The second igneous event. Above the lava sequence assigned to the Qordlortorssuaq Member, the volcanic sequence from the Vaigat Formation seems to consist of picrites and olivine-poor tholeiitic basalts, developed either in subaerial or subaqueous facies. The chemical composition of pillows (Table 1, no 2) and their lithology can be correlated with the Ordlingassoq Member on Disko. The subaqueous facies is described below.

An upper sequence of hyaloclastites extends for about 36 km along the Vaigat coast from just north of Nûk kitdleq in the west to point 1760 east of Pautût in the east. At Nûk kidtleq two thin hyaloclastite horizons, separated by 25 to 30 picrite flows, start in the otherwise subaerial lava plateau, and increase gradually in thickness towards the east until they merge. Thereafter they continue eastwards as a steadily thickening single horizon. Decimetre to metre thick deposits of mudstone with plant fossils occur at the base of the lower of these hyaloclastite horizons, and these mudstones, while still unconsolidated, were often deformed by advancing hyaloclastite beds. At Kugssinerssuaq the hyaloclastites overlie about 50 m of black shale assigned to the Naujât Member by Koch (1959). The hyaloclastites contain pillow fragments of picrite and olivine tholeiite, and they are regularly foreset bedded. The infilling pattern of the basin is quite complex in detail, but in general the hyaloclastite deposits have formed from subaerial lavas flowing into the water from the west and north-west. At the base of some breccia beds are found rounded beach pebbles of picrite and sometimes of metre-sized fragments of subaerial lava flows brought into the basin by advancing lava fronts.

On Disko, alkaline picrites and alkaline basalts form a separate unit (Manîtdlat Member) within the Ordlingassoq Member. Careful sampling of pillow fragments in the hyaloclastite from Tupaussat to Kugssinerssuaq has not revealed the presence of alkaline basic rocks in the area. If present at all, the Manîtdlat Member is restricted to the area between Nûk qiterdleq and Tupaussat.

Eastern Disko

The field work in eastern Disko was centred in the region around Kvandalen where the pre-existing information on the volcanic rocks is very sparse. Cretaceous and early Tertiary sediments occupy the lower levels up to 450 to 700 m a.s.l., and the interfingering and overlying volcanics belong to the Maligât Formation. Four areas in this region were investigated (fig. 3).

Marine shale. In the westernmost part of Kvandalen a recently retreated glacier has left an exposure below the base of the volcanic succession of about 20 m of dark grey silty micaceous shale. The shale contains plant debris including leaf imprints of needle and broad-leafed species. It also contains well-preserved mussels up to 1 cm in size which N. Noe-Nygaard (personal communication) has identified as *Nucula* sp. *Nucula* is a marine genus, and this indicates that the water basin was connected to the sea at least temporarily.

Hyaloclastites. A hitherto unknown sequence of hyaloclastites was discovered in the Kvandalen region (fig. 3). It consists of two units.

The lower unit is best developed in the westernmost part of Kvandalen where it is around 20 m thick. It consists of several metre-thick layers of relatively fine-clastic matrix-rich hyaloclastite with sparse pillows, rarely exceeding 5 cm in size, of plagioclase-porphyritic basalt. Trains of basalt pebbles and rounded shale xenoliths are seen at some levels. Centimetre-thick lenses and irregular layers of mudstone occur at two levels, one forming the top of the unit.

The upper unit is up to 90 m thick and consists of a regularly foreset-bedded coarse hyaloclastite locally

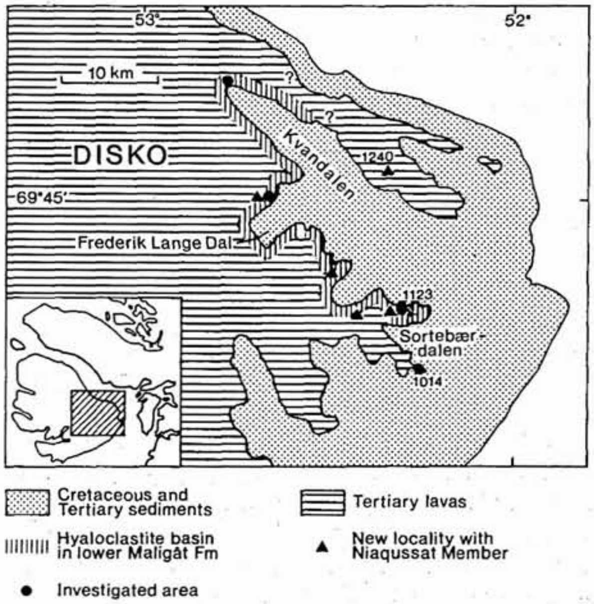

Fig. 3. Investigated area in eastern Disko.

rich in decimetre-sized pillows, covered by subaerial pahoehoe lavas belonging to the same eruptive unit as the hyaloclastite. The rock is a tholeiitic basalt (Table 1, no 3) with abundant microphenocrysts of platy olivine and phenocrysts of plagioclase. In Frederik Lange Dal two unusual composite lavas with basal parts of picrite and upper parts of olivine-poor basalt belong to the unit.

This hyaloclastite sequence forms an important marker horizon that has been observed over a large part of the Kvandalen region (fig. 3). It belongs to the lowest part of the Maligât Formation and demonstrates the existence of an easterly, water-filled sedimentary basin with a water depth of around 50 m. A considerable volume of this basin was completely filled in with hyaloclastite.

Subaerial lavas. After the infilling of the basin subaerial plateau basalt lavas became widespread. In the eastern part of Kvandalen and in Sortebærdalen quartzo-feldspathic sandstones occur between the lower lava flows, and a few coal layers up to several metres thick are present in some of these sandstones. Towards the west the sandstones diminish in thickness, and they are entirely absent in the westernmost part of Kvandalen. The major part of the subaerial lavas, as well as the underlying hyaloclastites, belong to the Rinks Dal Member of the Maligât Formation.

Nordfjord Member. Pedersen & Larsen (1987) reported

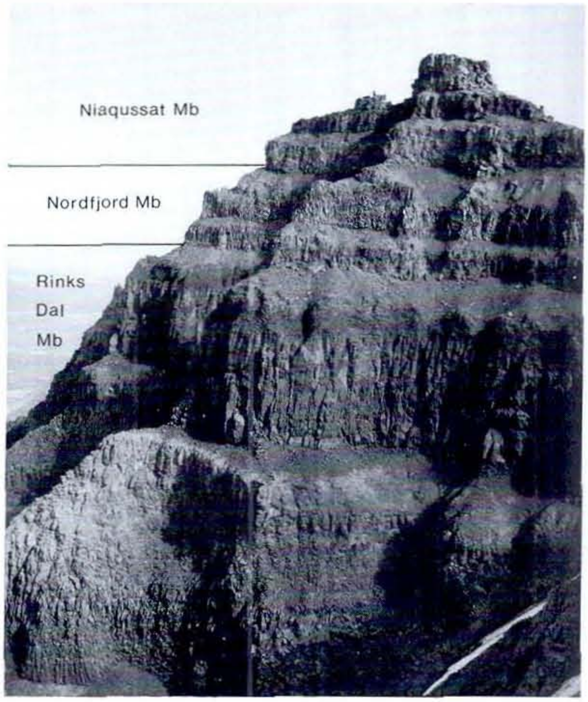

Fig. 4. Peak 1123 m in eastern Disko, seen from the west. Subaerial lavas of the Måligat Formation, with division into members as shown.

on the widespread occurrence of volcanic rocks from the Nordfjord Member, overlying the Rinks Dal Member, in the eastern parts of Nûgssuaq and Disko. This summer lava flows of silica-enriched basalt from this member were found to the south of Kvandalen, and it was established that a native iron-bearing lava from this member, known from the ridge to the north of Kvandalen, does not extend south of Kvandalen.

Niaqussat Member. Volcanic rocks from the Niaqussat Member overlie the Nordfjord Member and constitute the youngest volcanic rocks known from Disko (Pedersen, 1975). In western Disko widespread lava sequences belong to the Niaqussat Member; they include several iron-bearing lavas as well as iron-bearing craters and intrusions.

A characteristic feature of this member is the occurrence of silica-enriched olivine microporphyritic pahoehoe lavas (Table 1, no 4) which show flow-folding patterns and inhomogeneous vesicle distribution. Such lavas were identified among the uppermost lavas in Frederik Lange Dal, and they form the uppermost lava flows on the highest peaks along the south wall of Kvandalen (figs 3 and 4).

This discovery considerably extends the known range of the Niaqussat Member towards the east. It demonstrates that the volcanic sequence in eastern Disko, though thinned and degenerated relative to that in western Disko, still comprises the whole Maligât Formation. This is very valuable for the time-integrated basin analysis of the region.

Acknowledgement. The field work was supported by the Arctic Station in Godhavn.

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Geological photogrammetry using oblique aerial photographs

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Large areas in Greenland are covered by oblique aerial photographs, taken with 23 × 23 cm^2 photogrammetric cameras of the type 'Eagle' between 1948 and 1953. Though almost all of Greenland is now covered by modern vertical photographs, the oblique photographs still represent a valuable alternative source of geological information because of the perspective at right angles to the mountain side. So far, geologists have only been able to use the oblique photographs for visual interpretation.

New photogrammetric instruments like the Kern DSR 11/GP1 Analytical Plotter at the Institute of Surveying and Photogrammetry, the Technical University of Denmark, make it possible for geologists to map precisely from oblique photographs. This report describes an experiment where such photographs have been used for mapping of steep mountain sides for the production of geological maps, cross-sections and perspective views. Earlier results are reported by Dueholm & Garde (1986) and Heinesen (1987).

Experimental area

Current geological investigations along the south coast of Nûgssuaq and in northern and eastern Disko (Pedersen, 1985; Pedersen & Larsen, 1987; Larsen & Pedersen, 1988) concern Tertiary lavas and hyaloclastites, and their interaction with clastic sediments of non-volcanic origin. Volcanic units are traced from close to their eruption sites, where they originate as subaerial lava flows, to where they enter water-filled basins and develop into subaqueous lava flows and hyaloclastites which may show impressive foreset bedding. The volcanic units often encounter and interact with sandstones and mudstones from the Upper Atanikerdluk Formation (Koch, 1959). Accurate mapping of chronostratigraphic volcanic horizons from subaerial to subaqueous facies is important for integrated basin analysis.

The lithological units mapped are lava flows or sequences of lava flows, or hyaloclastite beds, which may be distinctive through variation in thickness, morphology or colour.

Photogrammetric orientation

Altogether four pairs of oblique photographs were analysed; the results from one stereo pair 514G1-NØ nos 8840 and 8841 from 1949 taken with an Eagle camera no. 15 is used as an example (fig. 1). The Eagle cameras used in the 1940s and 1950s for oblique photographs are not of the same quality as modern photogrammetric cameras. To increase precision it is therefore important not only to know the calibrated focal length of the camera, but also to obtain the calibration reports of the camera at the time of operation. Fortunately, the Geodetic Institute, Denmark, who took the photographs in 1949, can provide such reports for both cameras used in the experiment.

The Eagle camera has eight fiducial marks (réseau marks). Average residuals for an affine transformation of the measurements to the calibrated coordinates are about 25 microns. This is a relatively high value compared to the results obtained with modern cameras and film types, but it is not of great significance in the present experiment, and we were able to orientate three of four stereomodels and obtain satisfactory results. However, the fourth model, 514G1-NØ nos 8824 and 8825, taken with Eagle camera no. 15 on July 24 showed residuals on one fiducial mark close to 1 mm, which is a very high value, and the relative orientation also gave very poor results. The optical quality of these photographs is similar to other photographs taken with the Eagle camera 15 and is found to be of good geometrical quality, so the errors do not show up visually.

Errors during photography, or during storage, are likely sources of the large residuals. Thus, geologists should be aware during the planning stage that not all old photographs are of sufficient geometrical quality.

Control points needed for the absolute orientation of the stereopairs were digitized from topographic maps on the scale 1:250 000, and shore lines, intersecting rivers and mountain tops were identified and used.

The topographic maps of Nûgssuaq and northern Disko are of poor general quality and for this reason the identification of points is difficult. Therefore, the accuracy of the absolute orientation was low i.e. 30 m (1σ) in plane and 5 m (1σ) in elevation. Because of the oblique view the points were widely spaced (more than 30 km apart). Therefore, the standard errors on the scale in the orientated model were better than 1‰ in spite of the fairly large co-ordinate deviations. This was found to be satisfactory because the aim of the present experiment was relative measurements within a limited area.

The relative pointing precision varies over the model

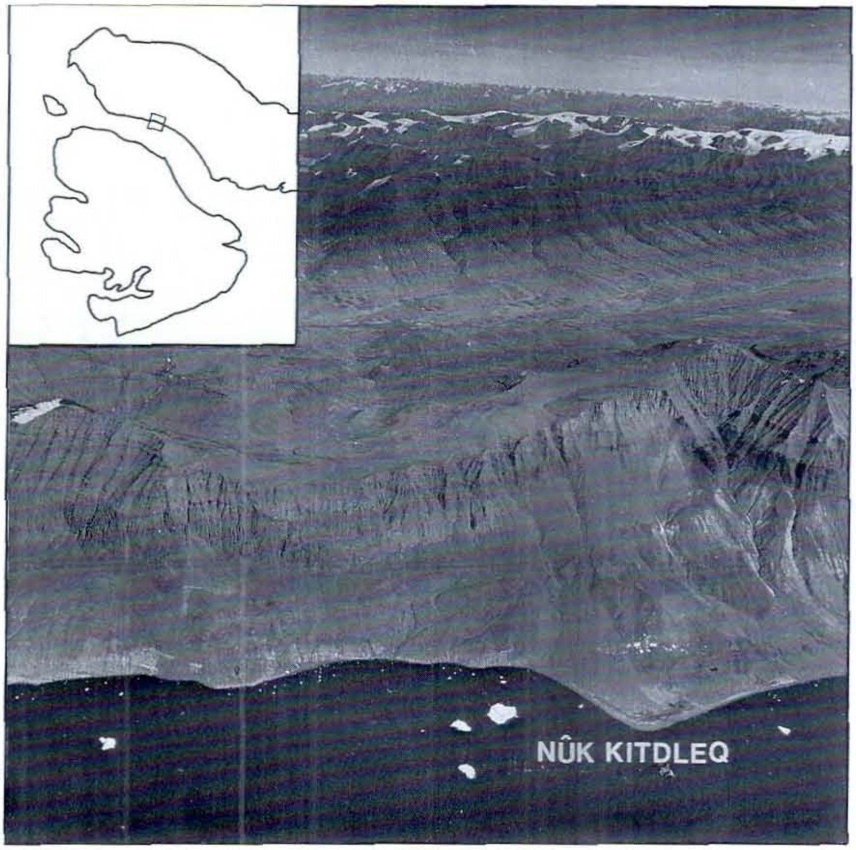

Fig. 1. Part of oblique photograph 514G1-NØ no 8840 from Nûk kitdleq at the south coast of Nûgssuaq. Courtesy Geodætisk Institut, Denmark.

because the oblique view gives a varying photographic scale. The mountain sides of interest were normally situated at a distance of about 7–8 km from the camera, resulting in a photographic scale of about 1:50 000.

As judged from the relative orientation, the photogrammetric precision obtained was about 20 microns, which gives a relative pointing error of about 1 metre at the scale 1:50 000. The resolution of the old photographs only allowed for a magnification of about five times, so that the photogrammetric interpreter observed the mountain side of interest at a scale of about 1:10 000.

During the project photograph overlaps of about 60 and 80% were tested. Although the mountain side of interest was close to orthogonal to the photograph direction, it was difficult to maintain the stereoscopic view at 60% overlap during the operation. Therefore, 80% overlap was used and is recommended for future work although the accuracy of measurement is about half that for 60% overlap.

Work procedure

The models were orientated by a photogrammetric operator, who took about two to three hours due to the troublesome work of finding and digitizing points from the topographic maps. Subsequently, contours were digitized by the photogrammetric operator, whereafter the geologist interpreted and digitized the relevant geological features in the model. Simultaneously with the data collection, a test plot was produced on an on-line plotting table. The average time spent by the geologist on a 10 km long profile was two days.

After the data collection, different views were plotted from the data-files. For this project a normal orthogonal projection, a section, and a perspective plot were drawn, and examples are given in figs 2 and 3.

Geological interpretation

The selected model, from Nûk kitdleq on the south coast of Nûgssuaq, was interpreted and measured photogrammetrically. Observations on the stereo model were supplemented by visual inspection of a series of colour slides taken in the model area from a helicopter. The features studied were the pre-Quaternary sedimentary and volcanic lithologies, with the main emphasis on the volcanic rocks from the Vaigat Formation. The three main lithologies are Cretaceous clastic sediments, Tertiary hyaloclastites and lava flows.

The exposures with Cretaceous sediments can be mapped easily, and several metre-thick units of sandstone or shale can be measured when they are not obscured by scree and landslides. Solifluction makes strike and dip measurements uncertain but it is easy to ascertain that sediments in the western part of the model show apparent dips of 6 to 9° towards the east, while sediments in the eastern part of the model are sub-horizontal.

The hyaloclastite basins are very suitable for this type of photogrammetry. Thickness of hyaloclastite units, and strike and dip of foresets can easily be measured and units with different colours can be mapped. The transition from subaerial lava facies to subaqueous hyaloclastite facies is easily recorded. As an example, lava flow *a* (fig. 2) developed into hyaloclastite foreset *a* with a strike of 01° and a dip of 27°E, and the water depth of the basin at the time of eruption of lava *a* was more than 491 m.

The varying erosion resistance and colours of the subaerial lava flows make it possible to map a sequence of lava flows and some individual lavas. Structures and variations in thickness of flows or series of flows as well as dips and strikes are easily measured. The method is not well suited for measuring and mapping dykes and faults cutting at oblique angles to the direction of observation.

The area of the model was subsequently visited in the field (Larsen & Pedersen, 1988) and the photo-interpreted features were found to be correctly recorded. However, some geologically important features which are not clearly distinguishable on the old photographs were also found.

Conclusion

During the project we found that the oblique view has the following advantages:

(1) The steep mountain sides facing the camera provide an excellent target for photo-interpretation.

(2) The angle of sight is suitable for some map projections and excellent for the construction of vertical sections.

(3) The angle of sight, which resembles that of the geologist's normal field situation, is of great help in the interpretation of the geology of sub-horizontal and dipping lithological sequences.

(4) In some models the oblique view permitted the identification of marker horizons over distances of up to 100 km, which may be very useful in reconnaissance work.

The following general problems were encountered with the use of oblique as opposed to vertical photographs:

(1) It is very difficult to base regional mapping on oblique photographs alone because of the varying scale and the shadowed areas.

(2) It takes more practice for the geologists to use the measuring mark in the oblique view than in vertical views.

The following special problems were encountered for the actual photographs taken with Eagle cameras around 1949:

(1) Photographs from one flight line were of poor geometrical quality.

(2) Photographs are of greatly varying quality for resolution and sharpness of details. Generally a magnification of more than five times brings out the grain in the photograph. Modern photographs would allow a much higher magnification.

However, despite these shortcomings, valuable geological information can be extracted from all the investigated photographs, though with variable geometrical accuracy and information density. In general, the methods are very suitable for producing detailed sections and profiles. Due to shadowed areas, however, regional mapping projects should still be based on vertical aerial photographs.

The variable quality of the old oblique photographs and the wide spacing on the flight lines raise the question of how to obtain new oblique photographs for geological investigations where, owing to financial and practical factors, new photographic missions with modern photogrammetric cameras cannot be expected. For this reason we have started experimental work on oblique photogrammetry with a series of colour diapositives taken with a Hasselblad camera from a helicopter. Several photographic series taken by F. Ulff-Møller for us in the Disko and Nûgssuaq areas in the summer of 1987 gave encouraging results.

Acknowledgements. The project is being carried out as a co-operative programme between the Institute of Surveying and Photogrammetry (ILF) at the Technical University of Denmark, the Geological Survey of Greenland, and the Geological Museum of Copenhagen. The project is supported by the Danish Natural Science Research Council and the Danish Technical Science Research Council.

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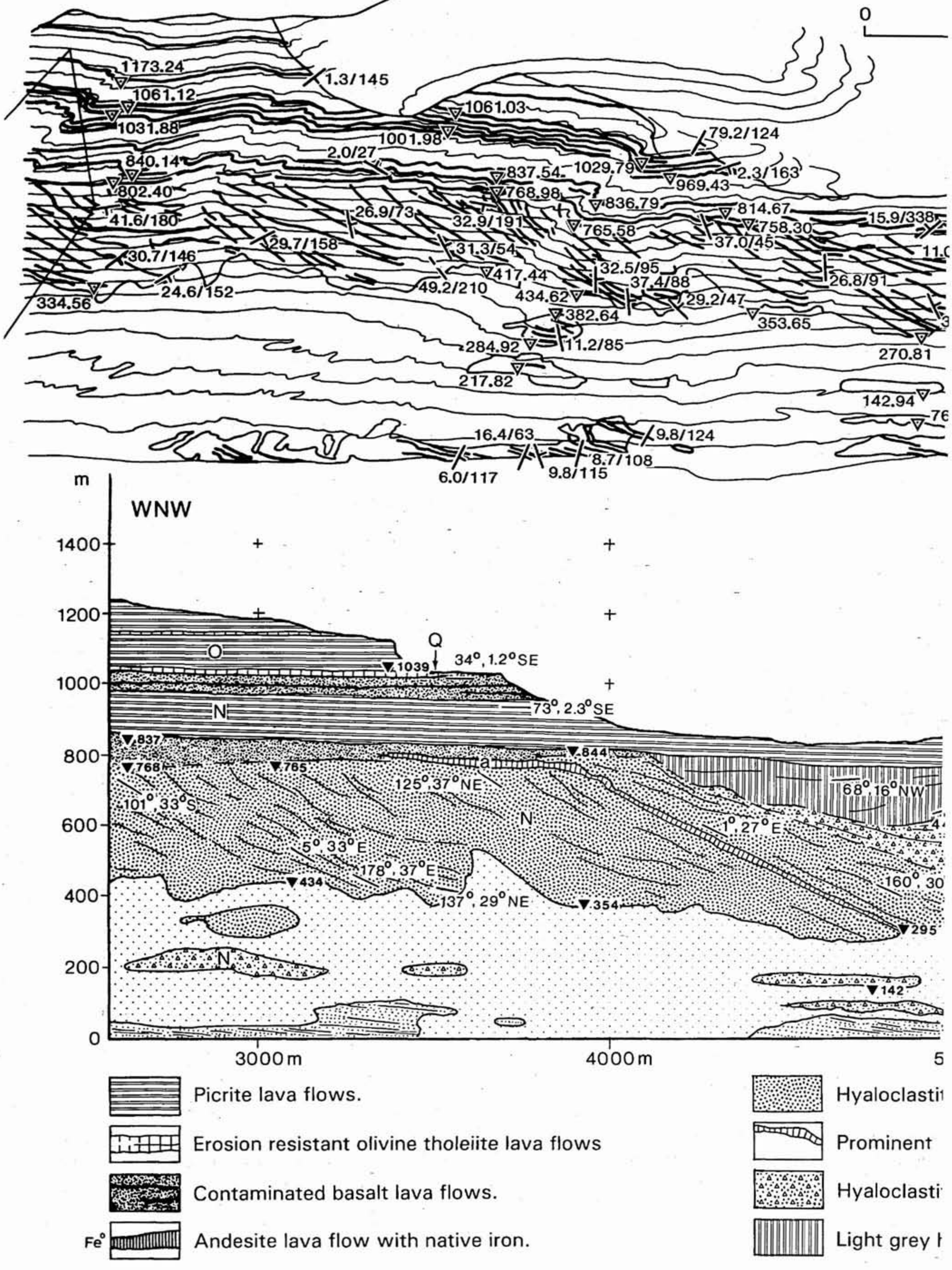

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2.0/27
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969.43
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802.40
768.98
836.79
814.67
41.6/180
26.9/73
32.9/191
765.58
758.30
15.9/338
37.0/45
29.7/158
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49.2/210
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29.2/47
382.64
353.65
284.92
11.2/85
270.81
217.82
142.94
16.4/63
9.8/124
8.7/108
6.0/117
9.8/115
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O
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73°, 2.3°SE
837
844
768
765
a
125°, 37°NE
68°, 16°NW
101°, 33°S
N
1°, 27° E
5°, 33°E
178°, 37°E
160°, 30
434
137°, 29°NE
354
295
N
142
3000 m
4000 m
Picrite lava flows.
Erosion resistant olivine tholeiite lava flows
Contaminated basalt lava flows.
Fe°
Andesite lava flow with native iron.
Hyaloclasti
Prominent
Hyaloclasti
Light grey

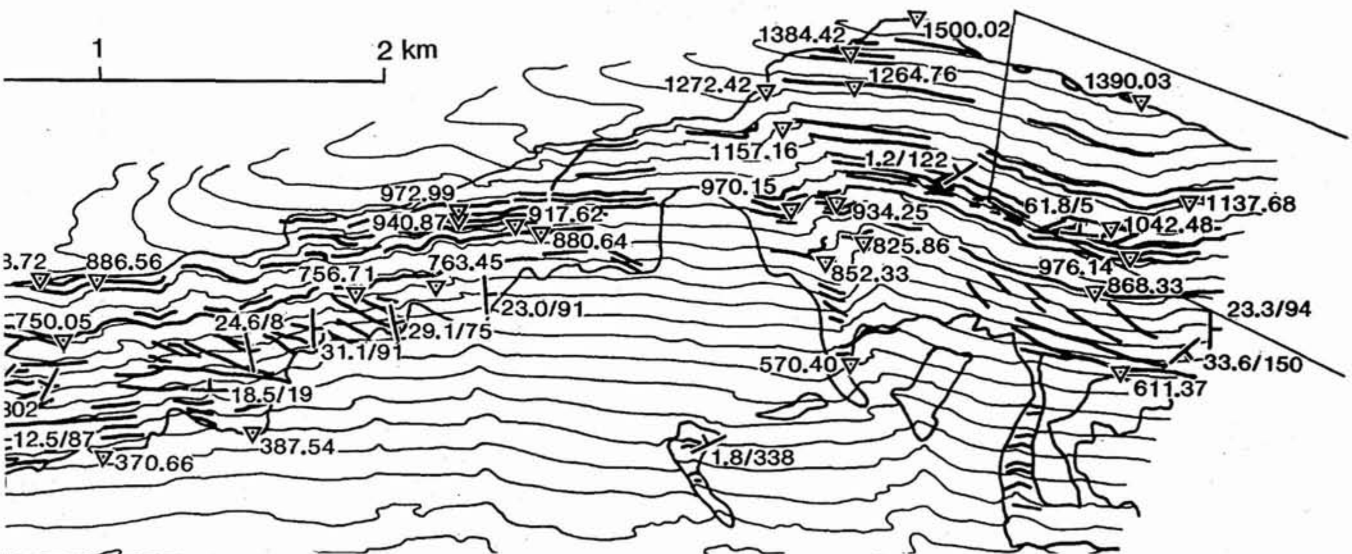

Fig. 3. Perspective view, as seen from SSW and an elevation of 20° at a distance of 10 000 m, plotted from data digitized from the photogrammetric model 514G1-NØ nos 8839 and 8840.

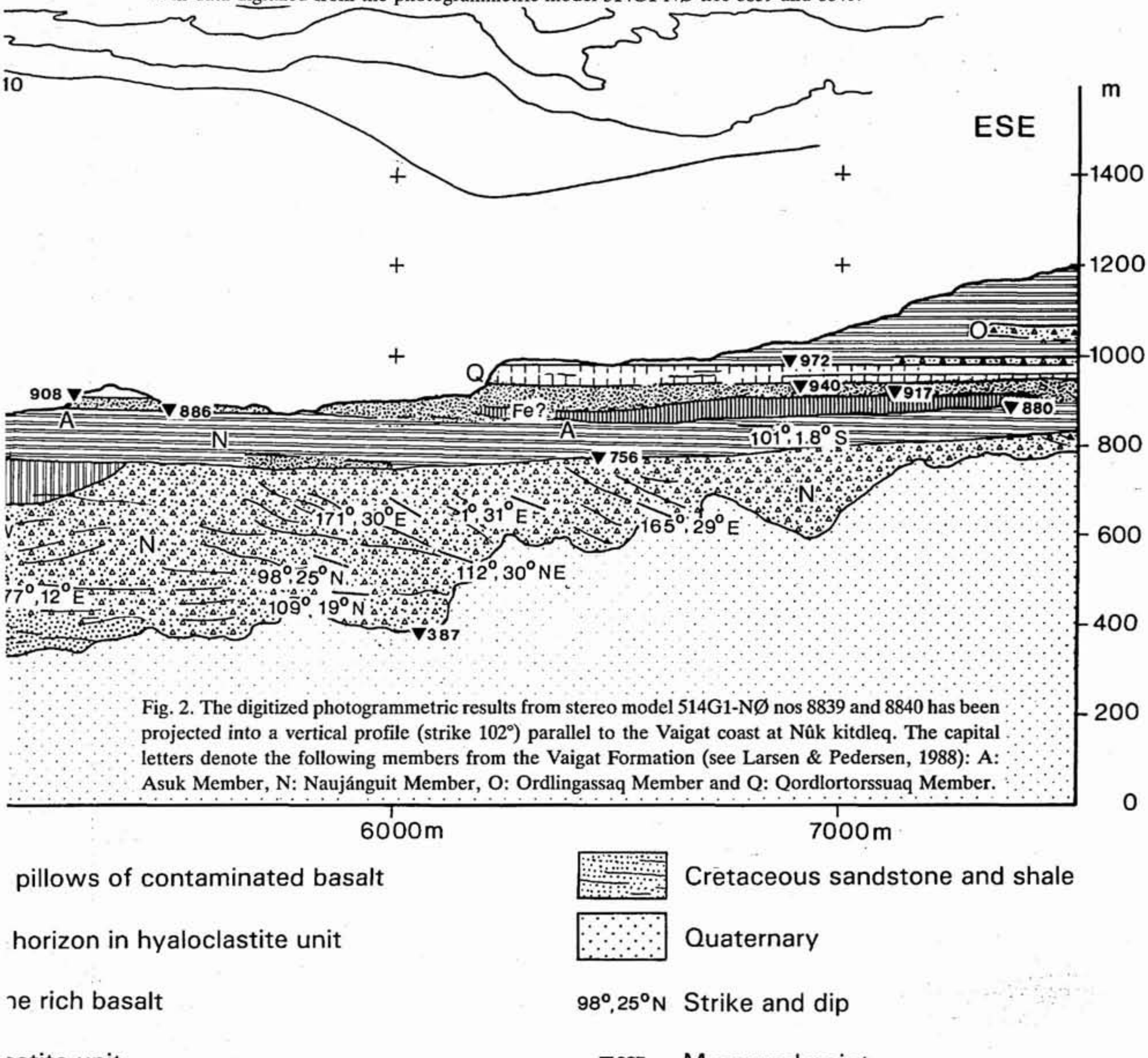

Fig. 2. The digitized photogrammetric results from stereo model 514G1-NØ nos 8839 and 8840 has been projected into a vertical profile (strike 102°) parallel to the Vaigat coast at Nûk kitdleq. The capital letters denote the following members from the Vaigat Formation (see Larsen & Pedersen, 1988): A: Asuk Member, N: Naujánguit Member, O: Ordlingassaq Member and Q: Qordlortorssuaq Member.

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Examples of bar accretion in fluvial sand, the Atane Formation, eastern Disko, West Greenland

Gunver Krarup Pedersen and Morten Willaing Jeppesen

The aim of the present paper is to supply additional sedimentological observations and to add new details to the existing interpretation of the Atane Formation on eastern Disko. Examples of epsilon cross-bedding reflect formation of point bars and indicate intermittent development of sinuous channels in the coarse-grained braided river. Large tabular sets of planar cross-bedding are interpreted as transverse bars and the coalescence of two such bars are discussed in detail. The field work was carried out during five days in July 1987 as part of a sedimentological research project supported by GGU and financed by SNF.

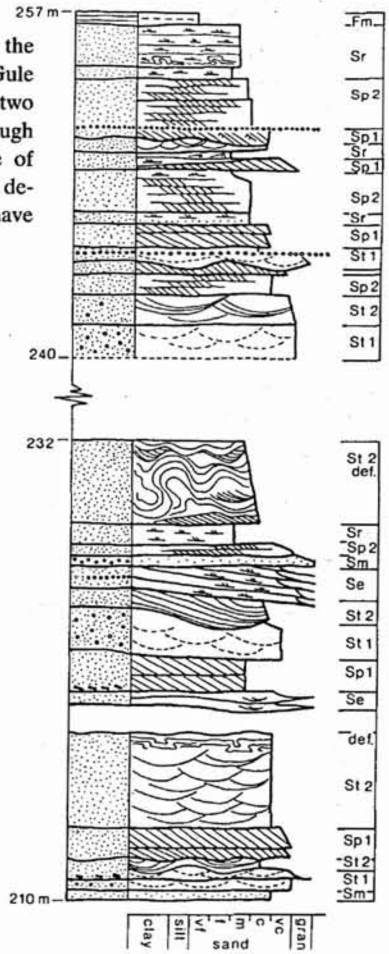
Background

White to pale yellow, slightly consolidated sand is widely exposed on eastern Disko below the Tertiary volcanic rocks. The sand was deposited in 10–40 m thick sequences capped by relatively thin clay horizons and eventually by coal seams. Palaeobotanical studies of these indicate a late Cretaceous age (Miner, 1932) and the sand is referred to the Atane Formation (Henderson *et al.*, 1976).

Johannessen & Nielsen (1982) and Bennike *et al.* (1981) studied the Atane Formation at Pingo and Skansen (fig. 1) and suggested deposition by a sandy braided river. They based their interpretation on the predominance of current-generated sedimentary structures, the unimodal palaeocurrents and the lack of channel abandonment or classical, fining-upwards point bar sequences characteristic of meandering rivers. The

Fig. 1. Location map, showing the distribution of the Cretaceous sediments in the Nûgssuaq Embayment and the localities where the fluvial sediments have been studied.

Fig. 2. A brief summary of the sedimentary facies of the Gule Ryg area accompanied by two vertical logs measured through exposures adjacent to those of figs 4 and 6. The lithofacies denominations of Miall (1977) have been used.



Facies	Description	Interpretation
Sm	Coarse-grained, poorly sorted structureless sand.	Rapid deposition from decellerating current. Possibly chanel floor (association with St1 at 210 m and pebble lag at 245 m).
St 1	Medium- to coarse--grained, locally pebbly sand with indistinct cross strata. Troughs are 10-20 cm deep and up to 100 cm wide.	Dunes migrating within channels or superimposed on bars or sandflats (Miall, 1977; Cant & Walker, 1976).
St 2	Medium-grained sand with distinct cross strata. Troughs are generally 8-10 cm deep and 30-50 cm wide.	
Sp 1	Solitary sets of planar tabular cross strata, often 50-150 cm thick. Facies 1, see text.	Transverse bar.
Sp 2	Medium- to very coarse-grained sand with tabular angular cross strata. Individual beds are 10-40 cm thick and form cosets up to 5-6 m thick.	Trains of straight crested sandwaves migrating across sandflats (Johannesen & Nielsen, 1982).
Se	Epsilon cross stratification. Facies 3, see text.	Point bar.
Sr	Cross laminated fine- to medium-grained sand, current generated structures.	Low energy subenvir., slough channels, top of sand flat, lower point bar.
S def	Soft sediment deformation ranging from oversteepened foresets to intricate slump folds.	Deformation of liquefied sediment caused by water escape (Plint, 1983).

palaeogeography of the northward flowing river is illustrated by Surlyk (1982).

Miall (1977, 1978) grouped braided rivers into six types of which the Platte type is characterized by an abundance of linguoid (transverse) bar and dune deposits, lack of well-developed cyclicity, and by a predominance of sand-sized sediment. The South Saskatchewan type is dominated by trough cross-bedded sand and characterized by thinning- and fining-upward cyclic sequences. Conceptual models of the distribution of bedforms and macroforms within these two types of rivers are illustrated as models 9 and 10 by Miall (1985). One or both of these models may encompass the Cretaceous sand of the Gule Ryg area.

Observations

Our observations show that vertical sequences are highly variable and fining-upward trends are weak and generally restricted to the upper 2–3 m below laterally persistent beds of silt, shale and thin coal seams (fig. 2). These horizons of fine-grained sediments delineate stages in vertical accretion of the fluvial system and range as third order bounding surfaces in the classification of Allen (1983) and Miall (1985). The sedimentary facies described by Johannessen & Nielsen (1982) are all recognized at Gule Ryg but occur in different proportions. A summary of the sedimentary facies is given in fig. 2, and only the transverse bar (facies 1 and 2) and the point bar (facies 3 and 4) are described below in greater detail and illustrated in figs 3, 4, 5 and 6.

Transverse bar

Facies 1: transverse bar front (figs 3, 4). Facies 1 comprises solitary tabular sets, (D) and (F), of large-scale cross-bedded, medium-grained sand with planar, tangential foresets. The foreset laminae are 1–2 cm thick, normally graded up-dip, while grading across individual foresets is insignificant. Large variations in grain size

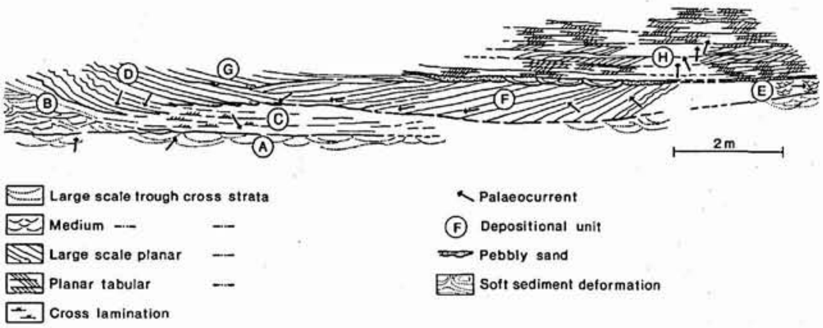


Fig. 3. Section through two transverse bars. Note the systematic change in palaeocurrent directions in F which indicate that the front of the bar was lobate.

are, however, found between successive foreset laminae. Some of these pinch out halfway down the slip-face and some are lenticular in cross section similar to the subaqueous grain flows discussed by Hunter (1985). The down-dip increase in grain size is especially pronounced in G, where the cross strata are seen to have sharp, locally lobate toes, which strongly point to deposition from avalanching. Coal debris is locally accumulated at the base of foresets. Reactivation surfaces are rare.

Topset strata are neither preserved in D, which is truncated upwards by a Quaternary erosion surface, nor in F which is cut by small scours and erosively overlain by a thin wedge of pebbly sand succeeded by cosets of tabular cross-bedded sand (fig. 3). Both D and F formed through lateral accretion on a core of trough cross-bedded sand which was an erosional remnant of slightly older deposits. Adjacent to the cores the foresets are oversteepened and slightly deformed by small slump folds (fig. 4).

Facies 2: bottom sets of the bar (figs 3, 4). Facies 2 comprises parallel-bedded, medium-grained sand with a low angle of dip denoted by C. The beds are *c.* 5 cm thick and form the continuation of the foreset laminae of D (fig. 4). Cross laminated intrabeds show ripple migration obliquely towards the slip face of the bar. This cross-lamination was produced by a subordinate current, similar to the low stage flow discussed by Collinson (1970) and Hobday *et al.* (1981).

Interpretation. Facies 1 is interpreted as avalanche deposition on the slip face of a bedform similar to the transverse, linguoid or cross-channel bars discussed by Collinson (1970), Smith (1971), Cant & Walker (1978), Miall (1977), Hobday *et al.* (1981). The systematic change in foreset orientation, indicating palaeocurrents from 310° to 240° in (F) (fig. 3), indicate that the bar front was lobate.

Figure 3 shows that the two bars F and D formed simultaneously. F migrated across the bottom sets C of D, but F was finally overlain by G which succeeded D (figs 3, 4). The coexistence of two bars support the interpretation of the bar front of F as lobate, because linguoid transverse bars are reported to occur in large fields in recent rivers (Collinson, 1970, Blodgett & Stanley, 1980).

The scarcity of reactivation surfaces may indicate that the discharge was relatively stable or that the bedform migrated more rapidly than fluctuations in discharge occurred. Oversteepened foresets and the small-scale slump deformations are compatible with rapid deposition during high-flow stage (Plint, 1983).

Reactivation surfaces were, however, generated where one transverse bar was overtaken by its successor (Johannessen & Nielsen, 1982, fig. 13e, f). Repetitive reactivation surfaces of this type were also seen at Gule Ryg, but appear here to be restricted to smaller bars (slip faces 50–60 cm high). A similar pattern of reactivation surfaces is described from mid-channel bars by Haszeldine (1983).

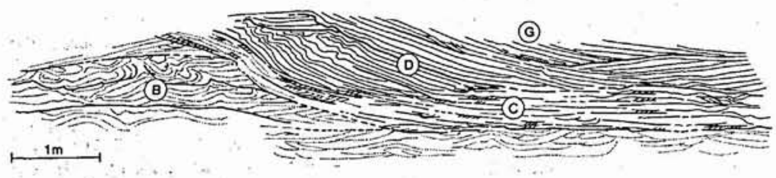


Fig. 4. Details of fig. 3 showing slip face strata D, bar bottomsets C and a core of older sediment B. The trough cross-bedding of B was distorted through water escape liquefaction prior to deposition of D. The foreset strata of D are oversteepened adjacent to B indicating continued plastic deformation. Same signature as fig. 3.

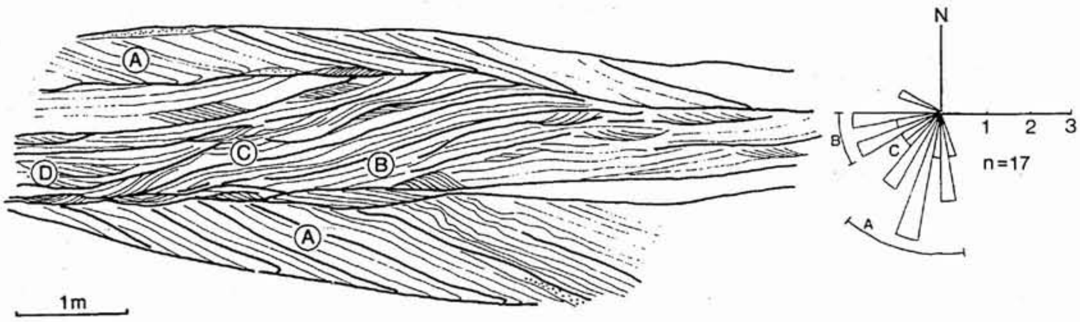


Fig. 5. Point bar intercalated between two transverse bars. A: transverse bars. B: epsilon cross-bedding. C: proximal part of facies 4. D: distal part of facies 4. Palaeocurrents of the depositional units are plotted for comparison.

Hobday *et al.* (1981) reported small scour channels on top of transverse bars and interpreted them as low stage dissection during partial emergence of the bar. Similar scours are seen in F (fig. 3) and in fig. 5.

Both bars are seen to have evolved through lateral accretion on cores of slightly older, trough cross-bedded sand (facies St2, units B and E, fig. 3). This supports the interpretation (Johannessen & Nielsen, 1982) of the river as characterized by frequent shifts between flooded or partially emergent bedforms.

Point bar

Facies 3: epsilon cross-bedding (figs 5, 6). Facies 3 consists of large-scale, gently dipping, planar cross-bedded, medium- to coarse-grained sand. The foreset laminae are 2–5 cm thick and normally graded across. Coal clasts of 1 cm are accumulated in the bottomsets. The unit contains several reactivation surfaces separating 15–20 cm thick bundles of foresets. The upper terminations of the foresets show offlap, while their lower terminations downlap on the channel floor or interfinger with facies 4. The top of the foresets dip *c.* 15° while the tangential bottom sets grade laterally into facies 4. In the upper part some small scour channels are seen.

Facies 3 is interpreted as epsilon cross-bedding. The lateral transition of facies 3 into facies 4 suggests a morphology like the lower point bar of McGowen & Garner (1970: fig. 4), though without associated chute bars. The properties of facies 3 are also comparable to those of the lateral accretion element or epsilon cross-bedding proposed by Miall (1985).

Facies 4: cross-lamination of lower point bar (figs 5, 6). The second depositional unit is built up by subhorizontal, 5–10 cm thick sets of cross-laminated medium-grained sand. The distal parts are comparable to the trough fill and foreset cross-stratification of McGowen & Garner (1970). In the proximal parts the sets become steeper, while the intrasets show ripple migration across the foresets of facies 3. Parallel laminae of sand are interbedded with the cross-laminated sand and drape the whole structure (fig. 5). The transitional zone between facies 3 and 4 is characterized by contortion of the cross-bedding and obliteration of all the internal small-scale structures. This feature is possibly related to liquefaction during the rapid deposition (Levey, 1978). Facies 4 was probably deposited in a lower point bar environment similar to the deposition at low stage flow conditions in the Amite and Colorado rivers (McGowen & Garner, 1970).

Interpretation. The point bars are thought to have formed under relatively quiet flow conditions in a 'secondary meandering river' between the emergent sand flats in the braided river system.

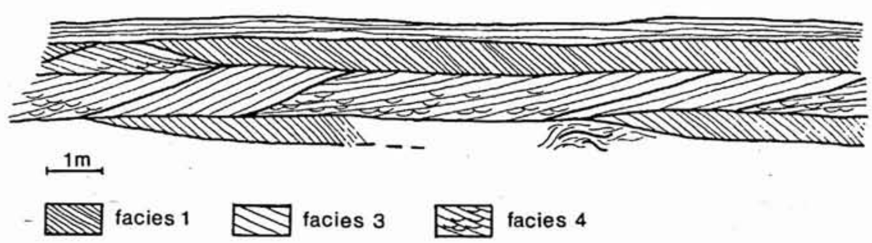


Fig. 6. Point bar complex showing that the lateral sequence from facies 3 to 4 shown in fig. 5 is repeated.

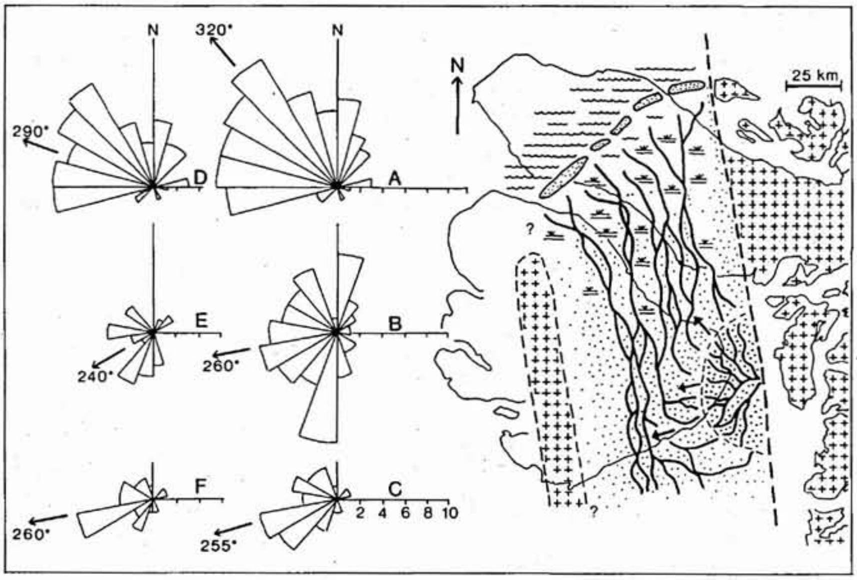


Fig. 7. Palaeogeographical sketch of the Nûgssuaq Embayment during deposition of the Atane Formation. Modified from Surlyk (1982). Palaeocurrent directions at Pingo, Gule Ryg and Skansen suggest sediment influx via an alluvial fan adjacent to the fault which borders the basin. Diagrams A, B, C include the total number of measurements, while D, E, F illustrate orientation of slip faces of facies 1. Diagrams A, C, D, F are redrawn from Bennike *et al.* (1981) and Johannessen & Nielsen (1982). A: n = 100, B: n = 73, C: n = 34, D: n = 79, E: n = 25, F: n = 27.

Figure 6 illustrates repetition of the lateral sequence composed of facies 3 and 4. The cyclic development of the point bar reflects varying discharge within this lower flood stage. Epsilon cross-strata represent maximum discharge and the cross-laminated and parallel laminated sand formed during minimum discharge conditions.

Discussion

The presence of epsilon cross-bedding associated with cross-lamination indicate deposition on point bars. Classical point bar sequences (Allen, 1970) with pronounced fining-upward trends and capped by overbank mud and coal are, however, lacking. Neither are channel abandonment sequences found. Point bar development was therefore presumably restricted to shorter periods of low-flow stage when sinuous channels developed between emergent sand flats. These may have been vegetated as comminuted plant debris is ubiquitous.

Transverse bars are seen in association with sand flat or point bar facies in figs 3 and 6. The former association corresponds to the facies model of the South Saskatchewan River (Cant & Walker, 1978). This model illustrates further the importance of channel floor deposition (facies St) which has occurred frequently in the sequence shown in fig. 2. The fluvial deposits in the Gule Ryg area do, however, have less pronounced fining-upward trends and less fine-grained sediment than has the South Saskatchewan model. Thin shale layers locally associated with thin coal seams are laterally continuous over hundreds of metres at Gule Ryg and divide the formation into 10–40 m thick sequences. Johannessen & Nielsen (1982) found that transverse bars predominate in the Pingo area and that signs of exposure are rare. They therefore concluded that the bars migrated in a wide, shallow river which only developed a braided channel pattern during low-flow stage.

Palaeocurrents

Palaeocurrent data from facies 1–4 are shown in detail in figs 3 and 5, and are included in fig. 7. The large spread in current directions is noteworthy, as is the lack of easterly directions. Most measurements were made of foreset orientations, and variations are found both between and within facies. This wide range of local flow directions indicates that the river was a multichannel system at least during low-water stages where the flow was channelled between metastable bars and sand flats.

The regional distribution of sedimentary facies within the Cretaceous of the Nûgssuaq Embayment shows fluvial sand on eastern Disko, delta-plain sand-shale-coal in southern Nûgssuaq and marine shales in northern Nûgssuaq. The main sediment transport was therefore towards the north as shown by Surlyk (1982).

Johannessen & Nielsen (1982) found unimodal palaeocurrents towards the north-west at Pingo while south-westerly directions predominate at Skansen (Bennike *et al.*, 1981). The current data from Gule Ryg are thus intermediate and all three current-roses are plotted for comparison in fig. 7. The many westerly current directions could reflect sediment transport into the Nûgssuaq Embayment via an alluvial fan adjacent to the regional fault which now separates this basin from the Precambrian terrain to the east.

Conclusion

Our observations support the existing interpretation of the Atane Formation on eastern Disko as a sandy braided river (Bennike *et al.*, 1981; Johannessen & Nielsen, 1982; Surlyk, 1982). The recognition of point bar sequences indicates the establishment of sinuous channels during low-flow stage. The palaeocurrents show a variety of directions, most with a westerly component. Compared with palaeocurrent directions from Pingo and Skansen this suggests sediment transport into the Nûgssuaq Embayment via alluvial fans along the eastern margin of the basin.

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Tourmaline in Precambrian supracrustal rocks from Aasiaat, central West Greenland

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Close to the town Aasiaat/Egedesminde supracrustal rocks outcrop on some small islands situated around the position 68°46′N and 52°38′W. The largest of these islands, Isuamiut, is barely a few kilometres long and the islands are situated more than 10 km from the mainland.

On Isuamiut two tourmaline-rich layers or tourmalinites have been found (Ellitsgaard-Rasmussen, 1954). Tourmalinite is a peculiar rock type consisting mainly of tourmaline together with quartz, feldspar or mica, sometimes with appreciable amounts of sulphides and/or scheelite. The chemical composition of the tourmaline in tourmalinites is to some extent indicative of the genesis of the boron. I therefore decided to collect some samples of tourmaline on Isuamiut when passing the area on a boat journey along the west coast of Greenland during July 1987.

Geology of the islands

The islands have been mapped in detail by Ellitsgaard-Rasmussen (1954). The following brief description of the geology is based on his work, but includes some further information gathered during the short visit to the islands.

The supracrustal rocks consist largely of metasediments and some metavolcanics which have been intruded by metagabbros. The metasediments comprise a surprisingly wide sedimentological set of rocks, ranging from very fine-grained black shales to coarse-grained metaconglomerates together with metamorphosed sandstones, quartzites and carbonate layers. The black shales which are found in layers up to tens of metres wide are rusty-weathering rocks sometimes stained with malachite and locally with appreciable amounts of graphite, pyrrhotite and small amounts of chalcopyrite. These layers can be traced for hundreds of metres along strike. The metamorphosed sandstones occur as metre-wide light grey to white layers with occasional clasts. The conglomerate layers are up to one metre wide with rounded to slightly angular pebbles which are up to 30 cm long. The carbonates occur as yellowish weathering layers up to one metre wide.

On the south-western part of Isuamiut a sequence of green mica schists, tens of metres thick with scattered grains of hornblende occur in which tourmaline-rich layers are seen (see below).

A thick sequence of staurolite-rich metasediments occurs on a small island south-east of Isuamiut. The staurolite crystals are up to 10 cm long and locally make up more than 50% of the rock. Staurolite has not been observed on the other islands.

Intercalated in the metasediments are frequent thin amphibolites which are locally quite rich in garnets. These amphibolites may represent basic tuffs.

The supracrustals have been repeatedly folded and metamorphosed under low- to medium-grade amphibolite facies. Subsequent to the deformation some carbonate and quartz veining has taken place.

Tourmalinites

Tourmalinites occur as layers up to a few tens of centimetres wide in the green mica schists of south-western Isuamiut. They are stratiform (Ellitsgaard-Rasmussen, 1954) and can be traced for some tens of metres along strike. The mica schists have up to 30% tourmaline. The tourmaline is black and often appears as clusters up to 5 cm long containing almost 100% tourmaline as fine crystals.

The tourmaline-bearing chlorite schists are cut by carbonate veins near which calcic amphiboles and thin massive tourmalinites are found. The carbonate veins have been deformed and at a later stage the tourmaline-rich schists have been cut by quartz veins. These quartz veins contain no tourmaline, but thin massive tourmalinites have been developed along their contacts.

On some foliation planes of the green mica schists very coarse-grained black tourmaline occurs. The tourmaline is up to 10 cm long sub- to euhedral crystals which have been broken up during deformation. These coarse tourmaline crystals and the tourmaline clusters are surrounded by a tourmaline-free halo of green mica schist.

In thin section the tourmaline is mostly brown to brownish green without any zonation. However, in some of the massive tourmalinite next to quartz and carbonate veins many of the tourmaline crystals display a well-developed zonality. These tourmalines consist of a light bluish centre surrounded by a brownish rim, and have sharp borders between the centres and rims.

Table 1. Microprobe analyses of homogeneous tourmalines

Sample No. (No. of analyses)	341214A (10)	341214 (11)	341215 (10)	341217 Centre (15)	Rim (15)
SiO_2	36.12	35.88	36.31	36.37	35.90
TiO_2	0.80	0.51	0.26	0.43	1.13
Al_2O_3	30.70	30.33	30.47	31.70	29.82
FeO	7.07	6.94	6.02	7.57	7.97
MgO	7.11	7.21	8.21	6.31	6.91
CaO	1.20	1.32	1.59	0.76	1.34
Na_2O	1.91	2.02	2.01	1.96	2.00

341214, 321214A, 341215 homogeneous tourmaline.
341217 zoned tourmaline.

All results in percent.

Chemistry of the tourmalines

Tourmalines from three samples of chlorite schist and one sample of massive tourmalinite next to a quartz vein have been analysed on a Jeol Super 733 microprobe using olivine, corundum, wollastonite and hematite as standards at 15 kV and 20 nA.

A total of 61 microprobe analyses were carried out where each analysis is an average of four analyses carried out 10 μapart. The results are listed in Table 1. The tourmalines have also been analysed for potassium, manganese and chromium, all of which appear in quantities at or below the detection limit.

The chemical composition of the tourmalines compares fairly well with that of tourmalines in tourmalinites from the Malene supracrustal rocks in the Nuuk area (Appel & Garde, 1987). The optical zonality observed in the tourmalines from the massive tourmalinites next to quartz veins is obviously due to significant differences in chemical composition between centres and rims of the tourmalines as shown in sample 341217 (Table 1). Sodium occurs in the same concentrations in the centres and rims, whereas MgO, CaO, TiO_2 and FeO are enriched in the rims of the zoned tourmalines and Al_2O_3 and SiO_2 show highest concentrations in the centres of the zoned tourmalines compared with the rims. These relationships correlate very well with the chemical zonation in zoned tourmalines from the Malene supracrustal rocks (Appel & Garde, 1987).

Conclusion

Until recently tourmaline was regarded as a mineral of granitic pedigree, but during the last decade it has become increasingly recognised that tourmaline can also be formed by other processes. In greenstone belts in North America and in Australia stratiform tourmalinites have been recognised and have been interpreted as products of submarine exhalative activity (Slack, 1982; Plimer, 1983). Recently similar stratiform tourmalinites have been recognised in the Malene supracrustal rocks of the Nuuk area, West Greenland (Appel, 1985; Appel & Garde, 1987) where they have been interpreted as exhalites which were precipitated on the sea floor contemporaneously with the deposition of the mafic tuffs which are frequently their hosts.

In his detailed description of the tourmaline-rich rocks in the supracrustals at Isuamiut, Ellitsgaard-Rasmussen (1954) interpreted the tourmalines as stratiform and suggested that the boron was derived from sea water and scavenged by precipitating clay minerals. His suggestion, which was quite controversial in the fifties, is in close agreement with the most recent theories regarding formation of tourmalinites.

The geological setting and the chemical composition of the Isuamiut tourmalines correspond very well with the composition of tourmalines found in metamorphosed pelitic sediments in the Archaean Malene supracrustal rocks (Appel & Garde, 1987).

The following mode of formation of the Isuamiut tourmaline-rich rocks is proposed. Boron, possibly of submarine exhalative origin, in sea water was scavenged by clay minerals. This first step is substantiated by the fact that certain clay minerals are able to scavenge appreciable amounts of boron, resulting in a pelite with up to 1000 ppm boron (Reynolds, 1965). After deposition of the boron-rich clay minerals at Isuamiut, tourmaline was formed during diagenesis at an early stage of metamorphism, resulting in fine-grained tourmaline crystals, more or less evenly distributed throughout the mica schist.

During prograde metamorphic conditions when the temperatures reached 400 to 500°C (Weisbrod, 1987) the solubility of tourmaline increased considerably, during which period the tourmaline-rich clusters and fairly coarse-grained tourmalines, which appear brownish in thin section, were formed. After the first deformation the tourmaline-rich mica schists were intruded by thin carbonate veins followed by thin quartz veins, the tourmaline then migrated at high temperatures to low pressure zones in the vicinity of the carbonate and quartz veins, and the bluish cores of the tourmalines were formed. The brownish rims around the bluish cores were formed at a later stage.

The presence of tourmaline-rich rocks and tourmalinites in supracrustal rocks has often been taken as an indication of economically interesting mineralisation such as the tourmalinites associated with the Broken

Hill ore body, Australia (Plimer, 1983). It seems therefore worthwhile to look closer at the supracrustal rocks at Isuamiut and at their conterparts on the mainland in order to locate interesting mineralisation such as massive sulphide ore bodies.

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Corundum crystals with blue-red colour zoning near Kangerdluarssuk, Sukkertoppen district, West Greenland

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Results are presented of field work at a corundum locality near the head of Kangerdluarssuk, east of Maniitsoq/Sukkertoppen, West Greenland. The authors carried out geological mapping and sampling over four days in August 1987 to evaluate the quality of the corundum as a possible gemstone.

Previous investigations

The corundum locality has probably been known to the local population for a long time. It was first visited by geologists in 1977 during reconnaissance mapping by GGU. The locality is mentioned by Secher *et al.* (1982), and by Petersen & Secher (1985) who reported rose red bipyramidal corundum crystals up to 15 cm long.

General geological setting

The Sukkertoppen district, which is part of the Archaean of southern West Greenland, has only been covered by reconnaissance geological mapping at a scale of 1:500 000 (Noe-Nygaard & Ramberg, 1961; Allaart *et al.,* 1978; Allaart, 1982). The area consists of granulite and upper amphibolite facies, quartzofeldspathic gneisses and supracrustal rocks, mostly amphibolites and their granulite facies equivalents, as well as clastic metasediments which are common in this part of the Archaean basement, e.g. on Hamborger Land north of Maniitsoq. Pods and lenses of ultrabasic rocks, sometimes several hundred metres long, occur within the supracrustal sequences and most commonly consist of ortho- and clinopyroxene and lesser olivine, together with Mg-rich amphibole and various low-grade alteration products.

Country rocks at the corundum locality

In the vicinity of Kangerdluarssuk there are several sequences of supracrustal amphibolites and metasediments with local ultrabasic lenses in the gneisses. The structural trend is persistently north-east with sub-vertical dips and a strong planar fabric particularly well developed in the metasediments.

The corundum locality occurs within one of these supracrustal sequences at the contacts between two small lensoid ultrabasic bodies and a spectacular horizon of bluish and rusty kyanite-garnet schist (fig. 1). The ultrabasic bodies are situated entirely within the metasediments which are in turn intercalated between calc-silicate banded and veined amphibolites. Pegmatites composed of quartz, feldspar, mica and occasionally black tourmaline occur within the amphibolites in the vicinity of, but not immediately adjacent to, the corundum locality.

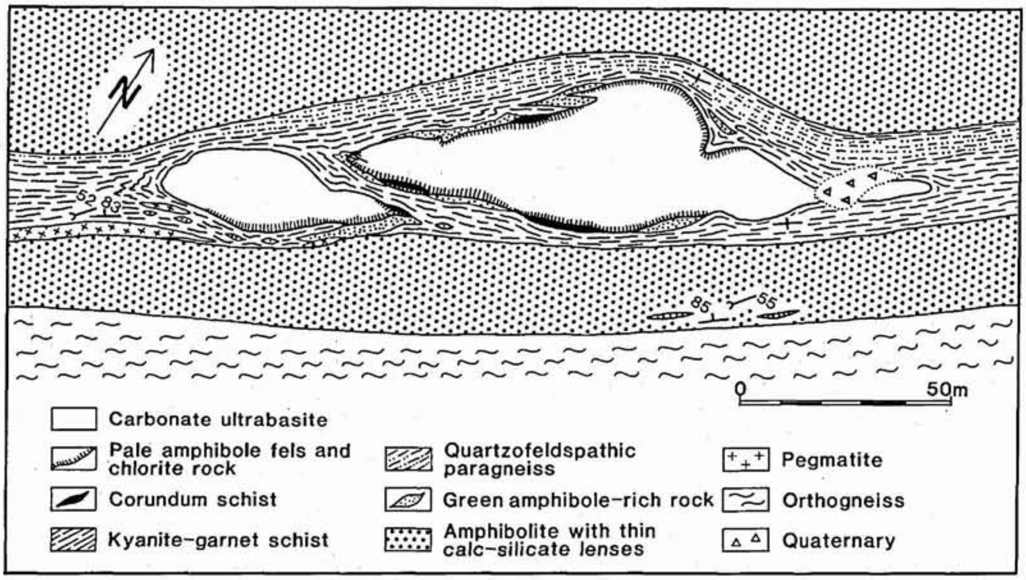

Fig. 1. Geological map of the corundum locality at Kangerdluarssuk.

The ultrabasic bodies are somewhat unusual for the West Greenland Archaean, being very olivine-rich and with appreciable amounts of interstitial carbonate. The outcrops are massive, dull grey on fresh surfaces, and have light brown weathering surfaces without the crumbling that is characteristic of most ultrabasic bodies in the region. Thin sections are dominated by subhedral fresh olivine grains up to *c.* 5 mm long with a weak preferred orientation. Interstitial carbonate, pale Mg-rich orthopyroxene and amphibole are common, as well as serpentine or Mg-rich chlorite, dark green spinel *sensu lato* (probably hercynite) and opaque oxide. Within the two ultrabasic bodies there are several veins up to *c.* 20 cm thick of slender prismatic to fibrous, pale brown to almost white, amphibole orientated perpendicular to vein surfaces.

The metasediments in the vicinity of the corundum locality both comprise kyanite-garnet schists and quartzofeldspathic paragneisses, with a gradual transition from one to the other over a few metres (fig. 1). The south-eastern part of the north-east striking metasedimentary unit consists of kyanite-garnet schists which are composed of alternating grey-brown quartz-feldspar-garnet rich layers a few centimetres thick, and bluish rusty layers of similar thickness with red garnet spots in a pale blue matrix. The blue matrix owes its colour to aggregates of prismatic kyanite *c.* 1–2 mm long, with additional fine-grained biotite, plagioclase, quartz, graphite, tourmaline, rutile and apatite. Fine needles in the matrix adjacent to kyanite aggregates may be kyanite or possibly sillimanite. The abundant light-pinkish-red garnets are up to 2 cm in size with local inclusions of staurolite. The grey quartzofeldspathic paragneisses at the north-western margin of the metasedimentary unit are uniform, weakly foliated, pale grey, fine-grained rocks consisting of quartz, plagioclase and microcline with minor biotite, muscovite, epidote and tourmaline.

The corundum occurrence

Between the two lensoid ultrabasic bodies and the kyanite-garnet schists characteristic metamorphic contact zones are developed which are at most a few metres, often only about one metre, thick. In the contact zone along the south-eastern side of the larger ultrabasic body there is a well-developed corundum-bearing schist. The innermost layer, along the margin of the ultrabasic body, is a pale yellowish to greenish amphibole rock with minor serpentine or Mg-rich chlorite, which is followed outwards by a coarse-grained greenish chlorite with or without amphibole rock with accessory graphite and rutile. Then follows a medium- to coarse-grained, brown phlogopite/biotite-amphibole-plagioclase rock with accessory apatite and zircon. Yellow-green apatite crystals up to 4 cm long were found in this layer.

The mica-amphibole-plagioclase rock just described grades into a corundum schist. This rock is composed of medium- to coarse-grained biotite, abundant barrel-shaped corundum porphyroblasts (about 10–20% of the rock), interstitial plagioclase and accessory graphite, rutile, tourmaline and zircon. Both graphite and rutile may be present in appreciable amounts. The corundum schist is about half a metre thick and extends for *c.* 20 m between the ultrabasic body and the kyanite-garnet schist. Adjacent to the corundum schists a brown staurolite-rich rock is locally developed. In this rock short prismatic staurolite prisms up to 1 cm in length, often with tourmaline inclusions, are set in a matrix of biotite, plagioclase, minor tourmaline and opaque oxides.

Corundum also occurs at three other, smaller outcrops at the margins of the ultrabasic bodies (fig. 1). Here the corundum crystals usually have a deeper red colour but are much smaller and not idiomorphic.

In several places at the contacts of the ultrabasic bodies there are lenses up to a few metres thick of green to almost black fine-grained rocks. The green lenses consist of graphite-rich, millimetre-banded, plagioclase-amphibole rocks, as well as massive monomineralic, dark green, fine-grained, amphibole rocks. Within the green rocks there are occasional small lenses a few centimetres thick, which consist of black spinel *sensu lato* (probably hercynite, as octahedra up to 2 cm in size), carbonate and chlorite.

The contact zones described above, like the ultrabasic bodies and the metasediments themselves, have been deformed during high grade *P-T* conditions. The contact zones and metasediments wrap around the ultrabasic bodies and are generally distinctly foliated, and tight irregular small folds are common in the contact zones especially at the tips of the ultrabasic bodies.

The corundum crystals

The corundum crystals are quite variable. At the largest occurrence south-east of the larger ultrabasic body described above, the corundums are usually idiomorphic barrel-shaped bipyramidal crystals, 2–5 cm in length with irregular terminations, very rough surfaces and a coarse horizontal striation. Several weathered-out crystals between 7 and 10 cm long were collected, as well as one broken specimen 16 cm long. The corundums are never transparent but are typically zoned with bluish grey cores and pale purple rims. Individuals that are purple throughout also occur. The corundum crystals are sometimes partially overgrown with graphite and/or rutile. The latter also forms inclusions up to about 1 mm in size perpendicular to the corundum *c*-axis.

When cleaned corundum fragments are viewed along the *c*-axis a silky lustre in a star-like fashion is sometimes visible. This lustre may be due to frequent tiny needle-like inclusions, probably of rutile, observed in thin section under high magnification. Test polishing has shown that weak asterism is present, but all the five polished specimens also contain many inclusions of larger (*c.* 0.1–1 mm) rutile grains.

Genesis of the corundum

At present it is only possible to present a general statement of the corundum genesis. Field relations and preliminary observations of thin sections indicate that the corundum schists were produced during the high-grade Archaean regional deformation and metamorphism. We think that the corundum was produced by metamorphic desilication reactions between the ultrabasic rocks and aluminous metasediments, and that chromium supplied from the ultrabasic rocks is responsible for the red colour of the corundum. The presence of both staurolite and kyanite in rocks that are considered to be cogenetic with the corundum schists indicates minimum *P-T* conditions of around 550°C and 5.5 kb (Winkler, 1979) during corundum growth.

Corundum of presumed metamorphic origin is known to occur elsewhere in schists adjacent to ultrabasic rocks. Schreyer *et al.* (1972), Ohnmacht (1974), and Cribb (1982) have described lensoid carbonate-orthopyroxenites (sagvandites) associated with amphibolites and kyanite schists from several localities in North Norway, where high-grade metamorphic corundum-bearing contact zones have developed. The above mentioned authors believe that the carbonate minerals (dolomite, magnesite and/or breunnerite, $Mg_9Fe(CO_3)_{10}$) in the sagvandites formed from metasomatic interaction between ultrabasic rocks and calcareous metasediments. The descriptions of the sagvandites and their contacts resemble the Kangerdluarssuk occurrence in many respects. It is possible that the widespread calc-silicate minerals in the amphibolites adjacent to the Kangerdluarssuk occurrence indicate a former source for the carbonate material in the ultrabasic lenses, and that decarbonation reactions have played a role in corundum-forming reactions.

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Border relations between the amphibolite facies Finnefjeld gneiss complex and granulite facies grey gneisses in the Fiskefjord area, southern West Greenland

Mogens Marker and Adam A. Garde

The authors carried out geological mapping in August 1987 at the south-eastern boundary of the Finnefjeld gneiss complex around 65°N. The field work was supported by the GGU cutter 'J. F. Johnstrup'.

Based on reconnaissance mapping in the 1950s Berthelsen (1951, 1957, 1962) divided the Archaean gneiss terrain in the southern Sukkertoppen district between Godthåbsfjord and Søndre Isortoq into three major tectonic units: the Nordland, the Finnefjeld and the Alángua complexes. This division was also followed by Noe-Nygaard & Ramberg (1961).

The Nordland complex of Berthelsen in the south (fig. 1) consists of mainly basic supracrustal rocks intruded by voluminous *c.* 3000 Ma old grey tonalitic gneisses, referred to as grey gneisses by Garde *et al.* (1986) and Garde (in press). The supracrustal rocks and orthogneisses show complex polyphase folding and have been metamorphosed under granulite facies conditions and later partially retrograded to amphibolite facies mineral parageneses.

The Finnefjeld complex in the central part of the region mainly consists of homogeneous grey biotite- and hornblende-bearing orthogneisses with amphibolite facies mineral assemblages. Berthelsen (1957, 1962) found occasional relics of hypersthene in the Finnefjeld gneiss complex and suggested that an episode of granulite facies metamorphisn preceded the amphibolite facies episode in the whole complex.

The Alángua complex to the north (fig. 1) consists of pelitic to semipelitic schists and basic supracrustal rocks embedded in homogeneous orthogneisses. The rocks are intensively folded and metamorphosed under high-grade conditions, with garnet and sillimanite in pelitic lithologies.

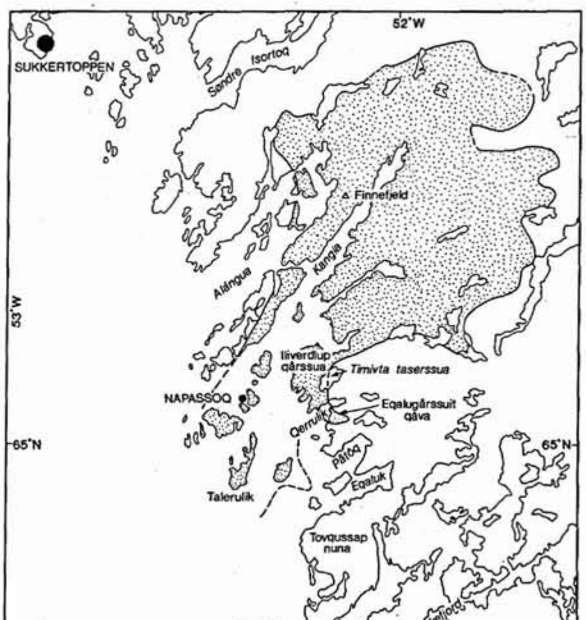

Fig. 1. Index map of the area between Sukkertoppen and Fiskefjord. The extent of the Finnefjeld gneiss complex is shown with dots.

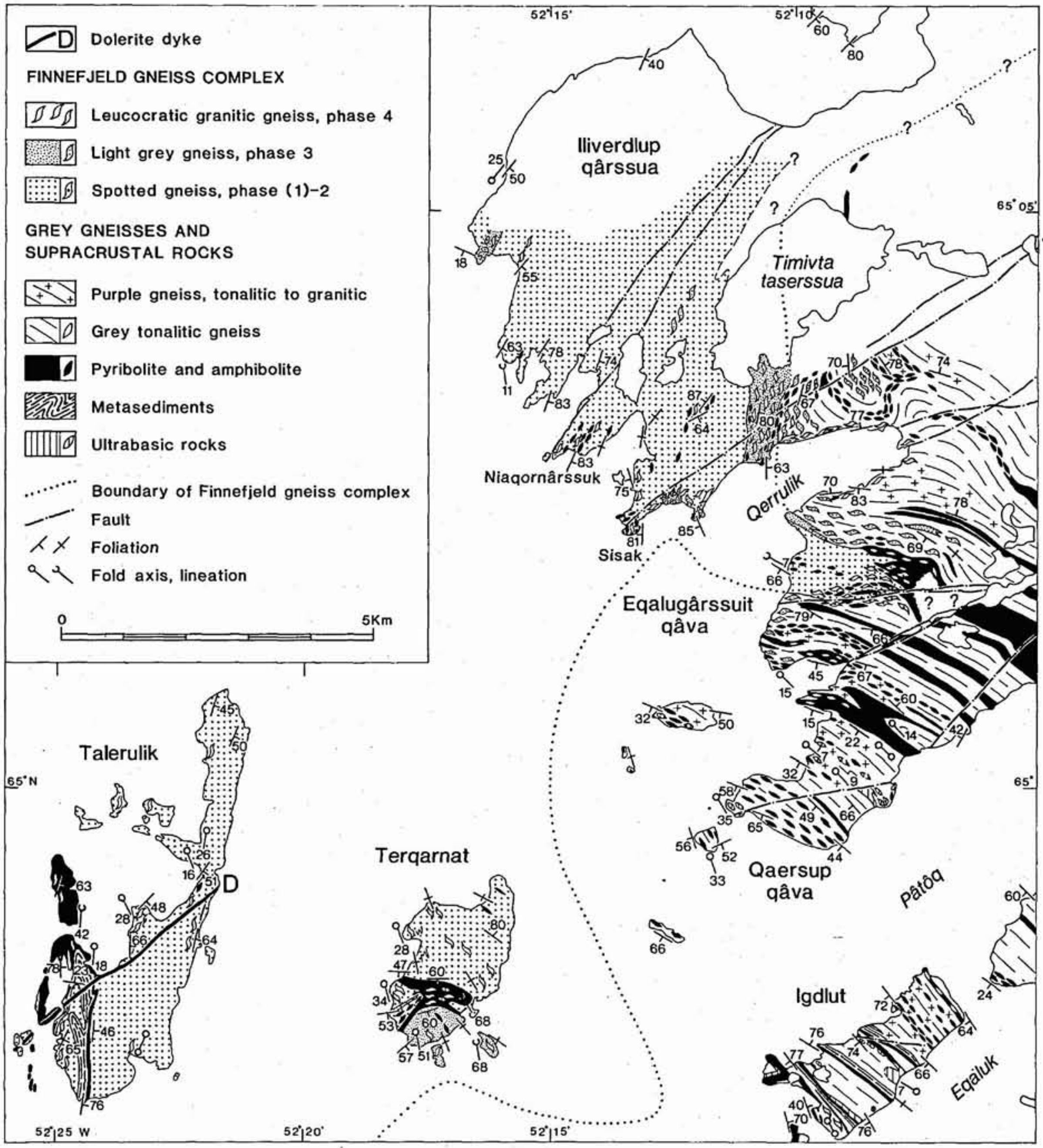

Fig. 2. Geological map of the southern border area of the Finnefjeld gneiss complex.

Berthelsen (1957, 1962) concluded that the Finnefjeld gneiss complex represents the retrograded and tectonised northern continuation of his Nordland complex which was thrust over the Alángua complex to the north-west. Berthelsen's investigations concentrated on the north-western part of the Finnefjeld complex, whereas no details of the border relations of its southern part were known until recently.

In connection with the systematic mapping for the Fiskefjord map sheet 64 V.1 N at scale 1:100 000 (Garde & McGregor, 1982; Garde, 1984, 1986), gneisses belonging to the Finnefjeld complex were reexamined in 1986 on the island of Talerulik in the north-western corner of the map sheet area (Garde *et al.*, 1987). This mapping revealed that the Finnefjeld gneiss complex is composed of at least four successive phases of homogeneous grey biotite (-hornblende) gneisses showing well preserved intrusive relations to each other. Similar polyphase intrusive relations were observed at a few localities on the coast of Iliverdlup qârssua.

The Finnefjeld gneisses show equilibrium amphibolite facies mineral assemblages and clear polyphase intrusive relations, and do not contain retrograde blebby textures (see e.g. McGregor *et al.*, 1986). Therefore the conclusion (Berthelsen, 1957, 1962) that the Finnefjeld gneiss complex represents retrograded gneisses belonging to the northern part of the Nordland complex, i.e. grey gneisses, could not be substantiated.

The boundary between the Finnefjeld gneiss complex and the grey gneisses was not located during reconnaissance mapping in 1986. However, the boundary did not appear to be in the position shown by Berthelsen (1962, Plate 1). Therefore the border area between these two important groups of gneisses was closely examined in 1987, in the light of the observations in 1986. The investigation concentrated on the area between Pâtôq in the south and Iliverdlup qârssua in the north, as well as the island of Talerulik.

Field relations and rock types

The mapping in 1987 revealed that the boundary between the Finnefjeld gneiss complex and grey gneisses is situated some kilometres further to the south-east (fig. 1) than hitherto believed (cf. Berthelsen, 1962 Plate 1; Allaart, 1982). The Finnefjeld gneisses were everywhere found to be intrusive into the grey Nûk-type gneisses (viz. the Nordland complex of Berthelsen). In the Iliverdlup qârssua area, the boundary curves around the western shore of Timivta taserssua (fig. 2). From the southern point of this lake it runs north–south to Qerrulik before it curves eastwards to Eqalugârssuit qâva, where it ends in an eastward-pointing apex. From the western coast of Eqalugârssuit qâva the boundary curves to the south (fig. 1) and it is inferred that it runs close to the southern tip of the island of Terqarnat (see below).

The best area to study the border relations is around Qerrulik and on Eqalugârssuit qâva. Particularly in the latter area, Finnefjeld gneisses are clearly seen to cut across rocks and structures within the grey gneiss terrain (fig. 2).

The grey gneisses and supracrustal rocks. In the area studied these are similar to those described from elsewhere in the Fiskefjord area (Berthelsen, 1960; Garde & McGregor, 1982; Garde, 1984, 1986; Garde *et al.*, 1987). The supracrustal rocks mainly consist of mafic, commonly hypersthene-bearing, rocks (pyribolites) with local metasediments, which are intruded by voluminous grey tonalitic gneisses. The gneisses are mainly light grey to grey rocks with retrogressive amphibolite facies parageneses and textures, and occasionally blebby textures visible in the field. Part of the gneisses, particularly around the head of Qerrulik, still contain a granulite facies mineralogy with hypersthene. Tracts of purple coloured gneisses (Berthelsen, 1960) with granodioritic to granitic compositions are widespread on Qaersup qâva and on the two small islands between Qaersup qâva and Terqarnat. It seems to be a general regional feature that purple gneisses are most common in a wide zone up to the north-western boundary of the grey gneisses. This pattern may suggest that it is genetically related to the Finnefjeld gneiss complex.

The Finnefjeld gneiss complex. This is composed of at least four intrusive phases. On the mainland in the Iliverdlup qârssua – Eqalugârssuit qâva area the predominant second phase (corresponding to Finnefjeld gneiss, Berthelsen, 1957, 1962) is a medium-grained homogeneous spotted grey gneiss of intermediate composition, often with feldspar porphyroblasts. In the less strained varieties biotite and/or hornblende occur within irregular 1–4 mm large, evenly distributed spots. The rock is massive and almost unfoliated on Eqalugârssuit qâva and north-west of Qerrulik in the southern part of the Finnefjeld gneiss area. It intrudes both grey tonalitic gneisses and pyribolite in the border zone to the south. At one locality original magmatic banding can still be observed (fig. 3). The strain increases towards the north-west, and northwards from around Niaqornârssuk the spotted, second phase gneiss is gradually transformed into a somewhat more fine-grained, foliated, rather striated rock. About 1 km east of Sisak the spotted phase is seen to form veins in an even older, dark grey, medium-grained gneiss (Finnefjeld gneiss phase one).

The third phase of Finnefjeld gneiss is a homogeneous, fine- to medium-grained, light grey, leucocratic rock with fine-grained dispersed biotite that forms a weak foliation. This phase, which may contain inclusions of the spotted phase two (fig. 4), seems to occur mainly in the border zone between the spotted (phase 1–2) Finnefjeld gneiss and the grey gneisses to the south (fig. 2). It intrudes the latter in a zone over one kilometre wide which forms a diffuse boundary to the grey gneisses. Their mutual relations are best observed on

Fig. 3. Compositional layering of presumed magmatic origin in Finnefjeld gneiss phase two; south side of Qerrulik.

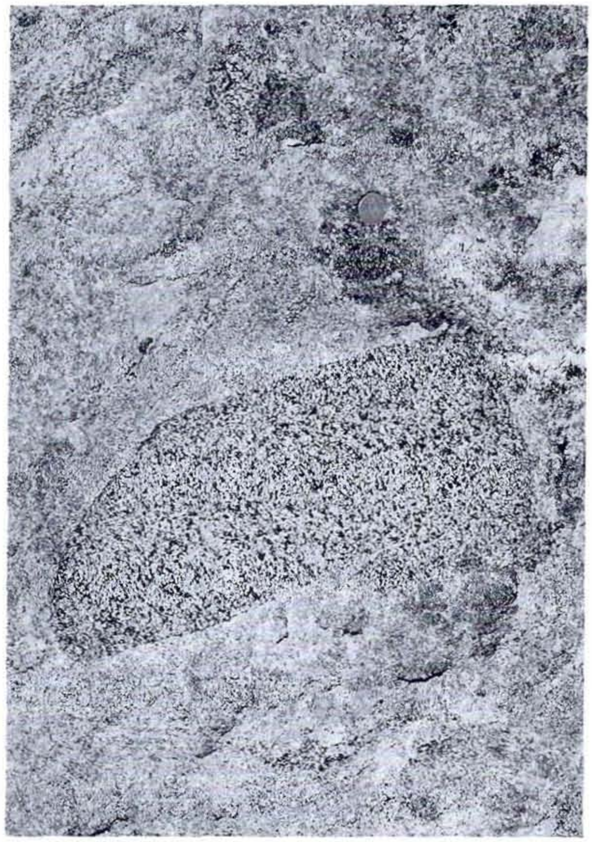

Fig. 4. Inclusion of spotted Finnefjeld gneiss (intrusive phase two) in Finnefjeld gneiss with dispersed biotie (intrusive phase three); c. 800 m north-east of Sisak.

clean-washed coastal exposures. The third phase of Finnefjeld gneiss also has a more pronounced foliation towards the north-west.

The fourth phase is a light greyish to white, medium- to rather coarse-grained, leucocratic, biotite gneiss or granite with granitic composition. It has high-angle intrusive contact relations to the earlier phases and to the grey gneisses in the border zone (fig. 2). The fourth phase seems to have been only partially involved in the folding affecting the first three phases of Finnefjeld gneiss, and it is therefore considered to be syntectonic. It has often been intruded parallel to axial surfaces of folds deforming the earlier intrusive phases. Phase four was succeeded by cross-cutting pegmatite dykes which may, however, be closely related to the fourth phase. Grey dykes seen at several localities cut all four intrusive phases of Finnefjeld gneiss.

The Finnefjeld gneisses on Talerulik (fig. 2) are mainly medium to light grey, very homogeneous, highly strained, biotite gneisses. Occasionally feldspar- and biotite-spotted textures like those common in the intrusive second phase of the Finnefjeld gneisses on the mainland can be discerned in the least deformed varieties. Cross-cutting and axial surface-parallel white granitic gneisses (intrusive phase four) are particularly common in the south-western part of the island, where large pegmatites also occur.

The island of Terqarnat (fig. 2) is divided by a curved, east–west trending amphibolite/pyribolite unit with an east-facing isoclinal fold. In the northern part of the island easily recognisable biotite (-hornblende) and plagioclase-spotted, amphibolite facies, Finnefjeld gneisses belonging to the first and second phases predominate, but all four intrusive phases are present. The area to the south of the amphibolite and on the two small islands to the south-east are composed of homogeneous, light grey, apparently un-retrograded amphibolite facies gneiss characterised by fine-grained dispersed biotite. This area is tentatively ascribed to the third phase of Finnefjeld gneiss in the border zone, but correlation with other islands and the mainland is uncertain.

Structural geology

A coherent structural analysis has not yet been made in the area studied here, or in the grey gneiss terrain further south, but some general statements can be made.

At least the north-western part of the Finnefjeld gneiss complex is highly deformed with high-strain fabrics and tight to isoclinal mesoscopic folds, although the structures are not always clearly mappable due to lack of marker horizons. The axial surfaces generally dip moderately to steeply to the south-east, parallel to the regional foliation. At Iliverdlup qârssua (fig. 2) the fold axes generally plunge south-west, while a few mineral lineations plunge gently to the SSE. At Talerulik most fold axes plunge gently to the south. At Terqarnat the plunge of folds is generally to the south-east and a strong NW–SE striking cleavage parallel to the axial surfaces is present. These folds everywhere affect the first three intrusive phases of Finnefjeld gneiss, and in part the fourth. The leucocratic fourth phase is often intruded as veins parallel to the axial surfaces, and folded veins of the fourth phase often show preferred orientations of their dark minerals (biotite, hornblende, garnet) in the same direction.

In the western part of Talerulik a large north–south trending unit of metasediments and (hypersthene-bearing) pyribolites, perhaps originally part of the Alángua complex, occurs adjacent to Finnefjeld gneiss. The supracrustal rocks were already folded before the intrusion of the Finnefjeld gneiss protoliths. During subsequent folding of both units, the supracrustal rocks were refolded into tight to isoclinal folds with subhorizontal north–south trending fold axes and steep east-dipping axial surfaces. The refolding was accompanied by recrystallisation and incipient anatexis in the metasediments under amphibolite facies conditions (Garde *et al.*, 1987). The synkinematic fourth phase of Finnefjeld gneiss intrudes the supracrustal rocks as well as the earlier phases of Finnefjeld gneiss.

Unfortunately the southern and western boundaries of the Finnefjeld gneiss complex are hidden beneath the sea. Along its south-eastern margin the Finnefjeld gneiss intrudes large-scale fold structures in the grey gneiss-pyribolite terrain between Qerrulik and Pâtôq (fig. 2). These folds in the grey gneiss terrain have horizontal to gentle SE-plunging axes and SW-dipping axial surfaces. The axial directions and the structural style suggest that these folds are equivalent to the Pâkitsoq fold phase on Tovqussap nunâ (Berthelsen, 1960, 1962), which took place under granulite facies conditions. If this correlation is correct, the intrusion of the Finnefjeld gneisses took place after the last main episode of folding in the grey gneisses (the Pâkitsoq fold phase). The possibility remains, however, that the folds in the grey gneisses between Qerrulik and Pâtôq are reorientated folds formed at an earlier stage than the Pâkitsoq phase, perhaps during the Smalledal fold phase of Berthelsen (1960). At present we prefer the former correlation; a safer conclusion may be reached when the fold interference patterns in the grey gneisses in the Fiskefjord area have been analysed.

Discussion and conclusions

Mapping of the Finnefjeld gneiss complex and the grey gneisses along its southern boundary does not support the conclusion (Berthelsen, 1957, 1962) that the former was reworked from the latter during retrograde metamorphism and tectonisation. That conclusion was influenced by the then topical 'granitisation theory' (A. Berthelsen, personal communication, 1987). Instead, we interpret the Finnefjeld gneiss complex as a batholith complex that intruded the grey gneiss-pyribolite terrain. The intrusion appears to have caused quite extensive remobilisation of the country rocks, giving rise to a hybrid border zone and back-veining of granitic material into the early phases of the Finnefjeld gneisses.

Mineral assemblages and rock textures strongly suggest that the Finnefjeld gneisses themselves never reached granulite facies conditions, although the Finnefjeld gneiss complex may contain older gneiss enclaves with a more complex metamorphic history. Berthelsen (1962) described some small patches with a dark variety of Finnefjeld gneiss that contain relic hypersthene close to the boundary to the Alángua complex. These patches may well represent partially digested xenoliths derived from the southern part of the Alángua complex, where hypersthene-bearing gneisses are known to occur (Berthelsen, 1962). Alternatively they may reflect local dry patches within the batholith (A. Berthelsen, personal communication, 1987).

It is evident from the new field work that the Finnefjeld gneisses intrude into earlier fold structures in the grey gneisses and supracrustal rocks. This may also be inferred from the 1:500 000 geological map Frederikshåb Isblink – Søndre Strømfjord (Allaart, 1982) which shows that large-scale structures in the grey gneisses terminate against the Finnefjeld gneiss. It is likely that the NW–SE orientated Pâkitsoq fold phase, which took place under granulite facies conditions in the grey gneisses, predates intrusion of the Finnefjeld gneiss protoliths. Folding of the Finnefjeld gneisses is possibly equivalent to the 'Posthumous phase' in Berthelsen's Nordland complex.

The following evolutional history for the Finnefjeld gneiss is proposed from our investigation. After the main deformation, granulite facies metamorphism and possibly retrogression of the grey gneiss terrain, the protoliths to the Finnefjeld gneisses were intruded as a polyphase batholithic complex. The earlier and more voluminous first and notably second Finnefjeld gneiss phases caused remobilisation of the country rock, giving rise to a hybrid border zone and back-veining into the early phases of the Finnefjeld gneiss complex. The light grey, dispersed biotite-bearing Finnefjeld gneiss (the third intrusive phase), which preferably occurs along the contact zone adjacent to the grey gneisses (fig. 2), seems to have formed during the remobilisation. The Finnefjeld gneiss complex was then deformed under amphibolite facies conditions in a NW–SE orientated stress field, during which folding and local overthrusting to the north-west took place (cf. Berthelsen, 1957, 1962). At the same time syntectonic granitic veins and minor bodies of the fourth Finnefjeld gneiss phase were intruded.

In several ways the Finnefjeld gneiss complex resembles the Taserssuaq tonalite some 75 km to the south-east (see Garde *et al.*, 1983, 1986; Nutman & Garde, in press). Both are large, polyphase, homogeneous batholithic masses of predominantly tonalitic gneisses with locally preserved igneous layering, which have intruded into already folded and metamorphosed supracrustal rocks and middle Archaean grey gneisses.

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Basic rocks of the inner Fiskefjord area, southern West Greenland

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During mapping for the Fiskefjord map sheet (64 V.1 N – 1:100 000) observations were made on the basic rocks. Earlier reports on the mapping of the Fiskefjord area have been given by Garde (1984, 1986) and Garde *et al.* (1987), and the reader is referred to these for more detailed descriptions of the general geology.

Most of the inner Fiskefjord area is made up of a variety of gneisses, correlated with the late Archaean Nûk gneisses. The most common type is a weakly banded, tonalitic gneiss, usually grey, sometimes with a lilac tint. Metagranites and more mafic gneisses of dioritic to quartz dioritic composition form mappable units as well as bands of inclusions in the tonalitic gneisses. The gneisses show intrusive relationships to a group of basic and ultrabasic rocks, mainly two-pyroxene basic granulites (named 'pyribolites' by Berthelsen, 1960) with minor dunite and norite. Small amounts of metasedimentary gneiss are associated with the norite.

Most of the gneisses have an amphibolite facies mineralogy, but with disequilibrium textures; these are thought to have arisen during retrogression from earlier granulite facies textures and mineralogy (Garde, in press). Unretrogressed patches, generally on a 10 m scale retaining granulite facies textures and minerals, are widespread in the gneisses; in contrast, the basic rocks are only locally retrogressed.

The gneisses and the basic rocks together underwent several phases of deformation. In the whole Fiskefjord area, three phases of folding followed by a phase of doming, have been recognised (Garde *et al.*, 1987); in the inner Fiskefjord area, two folding phases were distinguished. The last of these produced open to tight folds with steep, N–S trending axial planes, and variably plunging axes.

Norite and metasedimentary gneiss

The norite is a light grey, generally granular rock, rather friable and weathering to a white gravel, making it easily distinguishable in the field. It occurs mainly as a sheet within some of the pyribolites; the largest sheet has a maximum thickness of 150 m, though 10 to 50 m is more common, and it can be traced along the bottom of a pyribolite horizon for at least 15 km. In the central parts of thick sheets, the norite consists mainly of plagioclase and brown orthopyroxene (hypersthene), and is massive or weakly lineated. Towards the borders, amphibole replaces hypersthene and a prominent foliation may develop.

The norite is mostly homogeneous; at one locality hypersthene-rich layers were found, several metres thick. They are interpreted as an original igneous layering. Inclusions of ultrabasic rock are frequently found within the norite.

The norite often occurs together with garnet-bearing gneisses, interpreted as metasediments. They are rusty and friable, and appear deeply weathered in outcrop; besides garnet, they consist of varying amounts of quartz, plagioclase and biotite. Cordierite and sillimanite have not been seen in place; cordierite has been

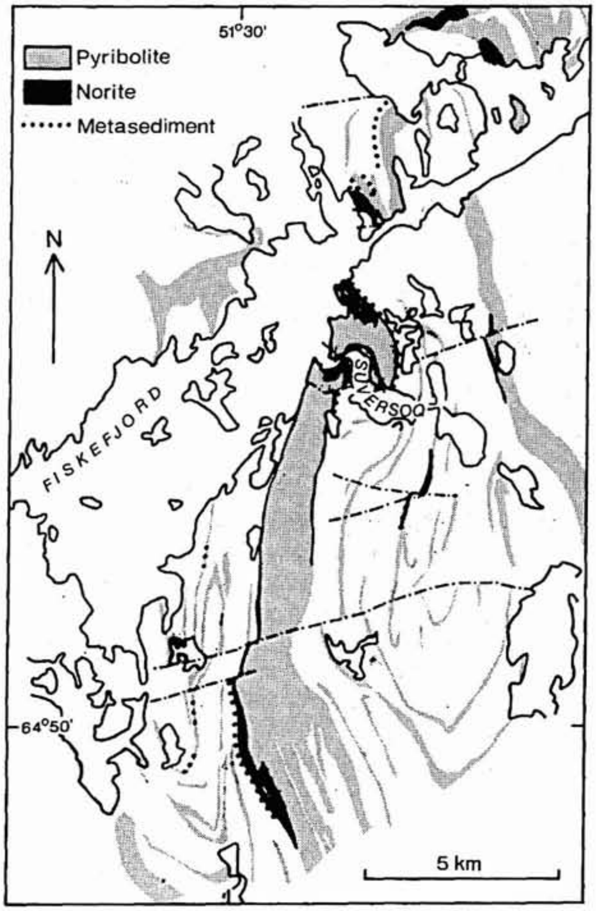

Fig. 1. Simplified map showing the distribution of pyribolite, norite and metasedimentary gneiss in the inner Fiskefjord area.

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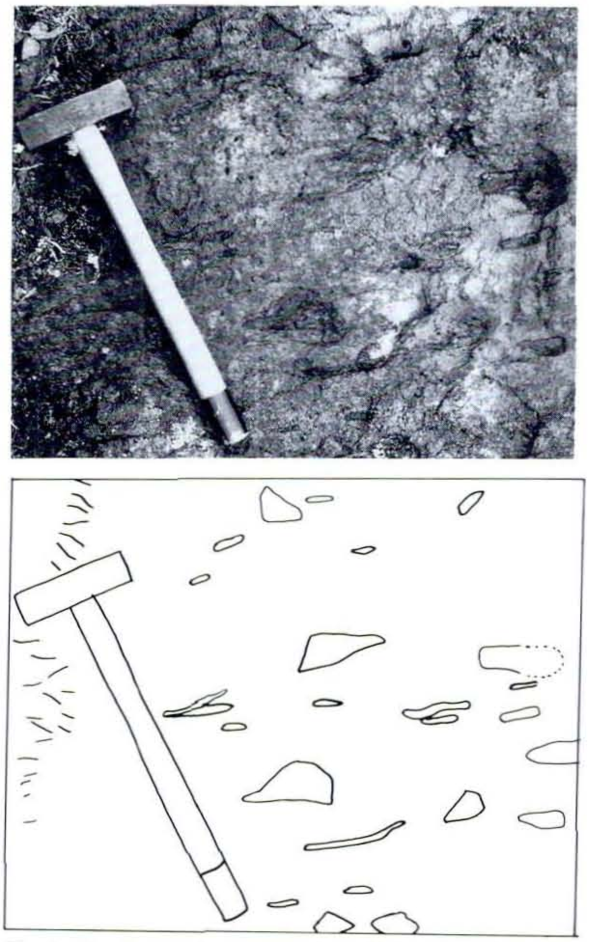

Fig. 2. Conglomeratic structure in pyribolite. Location: 2 km SSW of the mouth of Suversoq.

found in a loose block, probably from a nearby outcrop.

A compositional banding is generally present in the garnet-bearing gneisses. At one locality, this was seen to be cut by norite; the norite was undeformed along the contact, indicating that it had intruded the metasedimentary gneiss.

The norite – metasedimentary gneiss association is found in several of the pyribolites in the inner Fiskefjord area (fig. 1), and may eventually prove a useful marker horizon in the structural analysis of the area.

A conglomerate in the pyribolitic rocks

The pyribolites in the inner Fiskefjord area are generally rather homogeneous and structureless showing only an indistinct compositional banding; only in the large pyribolite at the mouth of Suversoq (fig. 1) are more variable types found. Here a 10–20 m thick horizon, with transitional boundaries to the surrounding homogeneous pyribolites, is rich in olivine and contains scattered millimetre-thick streaks of orthopyroxene. This horizon is situated 50–150 m above a large sheet of norite, associated with metasediment. Calc-silicate streaks occur in the surrounding pyribolite, testifying to its supracrustal origin.

At one locality the banded horizon exhibits a conglomeratic structure containing clasts of pyribolite. The clasts are mostly angular to subangular, oblong in section, up to 15 cm long and well separated by the surrounding matrix (fig. 2). They are bordered by a dark, 2–5 mm thick rim of orthopyroxene. The oblong clasts are parallel to the boundaries of the horizon. The clasts are completely recrystallised under granulite facies conditions, but appear to be undeformed.

If the whole conglomerate is indeed relatively undeformed, then the shape and orientation of its clasts suggest that it is a resedimented conglomerate, probably originating in an unstable, volcanic environment. However, the possibility of a flattened agglomerate or pillow breccia cannot be excluded.

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Scheelite in Malene supracrustals of the Ivisârtoq area, southern West Greenland

Peter W. Uitterdijk Appel

In 1982 the first traces of scheelite were found in the Godthåbsfjord area in heavy mineral concentrates from stream sediments. The same year the first *in situ* scheelite occurrences were found. In the following year a stream-sediment sampling programme was carried out in the area and in the area to the south. This programme demonstrated that scheelite is a frequent constituent of the 3800 m.y. old Isua supracrustal rocks and in the 3300 to 3000 m.y. old Malene supracrustals, whereas the gneisses in the Nuuk/Godthåb area are barren. The Malene supracrustal rocks, which form extensive outcrops, are the larger of the two supracrustal belts, and outcrops are found scattered over an area of at least 35 000 km^2.

During the field season of 1986 stream sediment samples were collected in the Ivisârtoq area of the inner Godthåbsfjord region, and they proved to contain high amounts of scheelite. It was thus decided to carry out field work with ultra-violet light, but early snow precipitation prevented any field work. During the field season the following year, another attempt was made to carry out field work with ultra-violet light in the Ivisârtoq area.

Geology of the Ivisârtoq area

The Ivisârtoq area has been mapped in detail by Chadwick (1986). The following brief account is based on his work combined with additional field information collected in 1987.

The Ivisârtoq area consists of Amîtsoq and Nûk gneisses, Malene supracrustal rocks, intrusive gabbro-anorthosites, pegmatites and a number of late dykes.

The Malene supracrustal rocks which form major outcrops in the area consist of a thick sequence of mafic volcanic rocks with thin intercalated sedimentary horizons together with thin ultramafic intrusives.

The Malene sequence at Ivisârtoq has been subdivided into a lower pillow-lava sequence, which is separated from an upper pillow-lava sequence by the so-called magnetite-bearing marker (see below). The upper pillow-lava sequence has been subdivided into several units which are separated by thin, but persistent sheets of ultramafic rocks of presumed intrusive origin (Chadwick, 1986). Intercalated in the mafic volcanics are frequent sulphide-rich metasedimentary horizons with disseminated and massive to semi-massive pyrite, which locally are fuchsite stained. These sulphide-rich zones are up to 50 m thick and can be traced for several kilometres along strike.

The magnetite-rich marker, which subdivides the volcanic pile of the Ivisârtoq area, is a sequence up to 50 m wide, which so far has been traced for well over 7 km. The sequence consists of a magnetite-bearing greenish schistose rock up to 10 m thick, which locally displays lapilli-like structures (Chadwick, 1986). The marker horizon furthermore includes white to light grey metasediments which are either metamorphosed sediments or acid volcanics. Very thin tourmaline layers occur in some of the metasedimentary beds. In the marker horizon there is also a horizon of skarn up to 20 m wide, which has been traced at intervals for well over 3 km. In this skarn horizon extensive scheelite occurrences have been found (see below). Malachite staining together with small amounts of sulphides such as pyrrhotite and chalcopyrite have been observed in the magnetic marker.

After deposition of the volcanic rocks and sediments the sequence has been repeatedly deformed, metamorphosed and intruded by numerous large pegmatite bodies. The pegmatites intruded at an early stage and have been strongly deformed. Rare examples of thin post-deformational pegmatites do however occur. The supracrustal rocks have been metamorphosed under amphibolite-facies conditions.

The supracrustals have been affected by extensive skarn alteration, whereby the amphibolites as well as the metasediments have been replaced by calc-silicates. In some places the amphibolites have been completely replaced, but usually they have been only partly replaced, with the central parts of pillows replaced and the outer parts of the pillows intact. Skarn is also frequently developed as clearly discordant patches and veins cross-cutting the amphibolites and metasediments.

There appear to be two distinct types of skarn, of which the most common occurs as patches of calc-silicates, while the less common type mainly occurs as distinct bands of garnet-rich calc-silicates. These garnet skarns are generally stratabound and scheelite bearing. The time of skarn formation is still uncertain, but it predates the deformation.

Stream sediment sampling

During the field season of 1987 only four stream sediment samples were collected in the Ivisârtoq area, but all of them contained high amounts of scheelite. The samples were collected in the same way as the previous samples from the Godthåbsfjord area (Appel, 1988), which involves sampling of 10 to 30 kg of gravel and sand which is sieved with a 1 mm mesh, and the fine material is then concentrated by panning. In the final heavy mineral concentrates the number of scheelite grains are counted after drying the sample. One of the collected samples contained 80 grains of scheelite per litre of fines, while the other samples contained well over 500 grains of scheelite per litre. These scheelite anomalies are the highest so far recorded in the Godthåbsfjord area (Appel, 1988). It is nevertheless quite clear that none of the scheelite-rich horizons located in the Ivisârtoq area (see below) would have been found by stream-sediment sampling alone. The drainage pattern in this area is poorly developed, and the streams generally carry only small amounts of sediment.

Scheelite occurrences

Field work with ultra-violet light was carried out for about three weeks in the Ivisârtoq area, involving an almost complete traverse across the whole supracrustal belt, and also detailed work in specific areas.

Scheelite is widely distributed throughout the lower as well as the upper sequence of pillow lavas where it occurs as small disseminated grains, but it is generally in small amounts. Scheelite is often found to be particularly abundant in the central part of pillows which have been wholly or partly replaced by skarn. In such pillows the scheelite mostly occurs in the central part of the pillows. Scheelite is also often seen in skarn patches replacing the amphibolites.

In the amphibolites, massive scheelite is found locally in clearly discordant veins which can be traced for a few tens of metres along strike. Scheelite is furthermore found as porphyroblasts up to 10 cm long and as joint coatings in the amphibolites. It is apparently equally abundant in the lower and upper sequence of metamorphosed pillow lavas.

From an economic point of view the most interesting scheelite occurrences are found in garnet-bearing skarn horizons. One scheelite-rich garnet skarn has been found interlayered in the lower amphibolite sequence. This horizon is about 1 m wide and was traced for about 50 m along strike.

This horizon lies juxtaposed to the magnetite-bearing schistose rock in the marker horizon. A skarn horizon up to 20 m wide, which has been traced for well over 3 km along strike, is garnetiferous and hosts extensive disseminated scheelite yielding a high grade. The garnet skarn horizon contains bands up to 4 m thick with high amounts of scheelite. The average width of the well-mineralised garnet skarn is of the order of one to two metres. The scheelite in these 'high grade' bands is mostly medium to rather coarse grained, and displays a white to light yellowish fluorescence indicating a molybdenum content of the order of 1%.

There was inadequate time for a systematic chip-sampling programme so no estimates are be presented on the grade of this 'high grade' zone. However, analysis of 12 grab samples yielded an average of 0.97% tungsten with a range of 0.15 to 1.91% (Appel, 1988).

Discussion

A detailed discussion of the genesis of the scheelite in the Ivisârtoq area is premature. In the Malene supracrustal rocks of the outer Godthåbsfjord area quite extensive stratabound scheelite mineralisation has been found in banded amphibolites and in tourmalinites (Appel & Garde, 1987), and it has been shown that this scheelite mineralisation is of submarine exhalative origin and that the scheelite precursor was precipitated on the sea floor contemporaneously with the deposition of the mafic tuffs and the chemical precipitation of the tourmaline precursor in the tourmalinites (Appel & Garde, 1987).

In the Ivisârtoq area the scheelite is generally stratabound and the highest scheelite concentrations occur at the same stratigraphic level as the tourmaline-banded metasediments. It is thus realistic to invoke the same origin for the scheelite occurrences in the Malene supracrustals at Ivisârtoq as in the other parts of the Malene supracrustal belt.

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Evolution and emplacement of Archaean terranes in the Kapisigdlit area, southern West Greenland

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In the Godthåbsfjord region, early Archaean Amîtsoq gneisses are found in association with the middle Archaean Nûk gneisses (McGregor, 1973). Early models of crustal evolution in the Godthåbsfjord region suggested that the Amîtsoq gneisses were reworked in a major crustal accretion-differentiation event, which was marked by the injection of the voluminous Nûk gneisses (e.g. Bridgwater *et al.*, 1974; McGregor, 1979; Moorbath *et al.*, 1986). This event has been interpreted as having culminated in a granulite-amphibolite facies metamorphic event at *c.* 2800 Ma, which outlasted all significant ductile deformation (e.g. Wells, 1979; Coe, 1980).

However, recent detailed mapping combined with U-Pb zircon dating (Friend *et al.*, 1987; H. Baadsgaard, personal communication, 1987; P. D. Kinny, personal communication, 1987) shows that between outer Ameralik and Tre Brødre 40 to 75 km to the south and west of the Kapisigdlit area (fig. 1), there are three lithologically distinct terranes. These terranes were tectonically juxtaposed and then folded together under amphibolite facies conditions between *c.* 2750 and 2550 Ma. The *Tasiusarsuaq terrane* is dominated by middle Archaean gneisses affected by *c.* 2800 Ma granulite facies metamorphism. The Tasiusarsuaq terrane structurally overlies the *Tre Brødre terrane*, which is dominated by a suite of 2800–2750 Ma granodioritic gneisses, named Ikátoq gneisses (Nutman & Friend, in press). The Tre Brødre terrane in turn structurally overlies the *Færingehavn terrane*, dominated by the > 3600 Ma Amîtsoq gneisses. Unlike the overlying Tasiusarsuaq terrane, neither the Tre Brødre terrane nor the Færingehavn terrane underwent granulite facies metamorphism at *c.* 2800 Ma.

The main aim of the field work in 1987 was to see if the terranes and structural history established between outer Ameralik and Tre Brødre could be traced to the north and east into the Kapisigdlit area (fig. 1). An invaluable starting point of this work was an unpublished 1:100 000 geological map of much of the area produced from helicopter reconnaissance mapping in 1976 (Allaart *et al.*, 1977).

Terranes of the Kapisigdlit area

Færingehavn terrane. Units of streaky, tonalitic-granodioritic gneisses, with closely-spaced irregular pegmatite banding outcrop in the Kapisigdlit area. These gneisses are cut by abundant, locally discordant amphibolite dykes, correlated with the Ameralik dykes (McGregor, 1973), and are continuous with the type Amîtsoq gneisses of the Færingehavn terrane in the outer Ameralik area. However, no units of augen granite gneiss, a common Amîtsoq lithology in the outer Ameralik region (e.g. Nutman *et al.*, 1984), were found. The streaky Amîtsoq gneisses in the Kapisigdlit area contain inclusions of banded amphibolite, banded iron formation, and clinopyroxene-rich rocks. Collectively, these resemble the Akilia (supracrustal) association found as inclusions in the Amîtsoq gneisses of the outer Ameralik area (McGregor & Mason, 1977). The Amîtsoq gneisses are not intruded by the late Archaean Ikátoq gneisses which are the dominant lithology in the adjacent Tre Brødre terrane.

Amîtsoq gneisses outcrop on the north and south shores of inner Ameralik, Itivdleq and Ameragdla. However, they were not found to be as extensive as was stated by Allaart *et al.* (1977). Coastal work and helicopter reconnaissance failed to find either the large unit of Amîtsoq gneisses running from inner Ameragdla eastwards along the valley of Austmannadalen, or the Amîtsoq gneisses east of Kapisigdlit, both described by Allaart *et al.* (1977).

Tre Brødre terrane. Much of the Kapisigdlit area consists of rather homogeneous biotite ± garnet granodioritic gneisses, with widely spaced pegmatite layering. These gneisses grade locally into rather more schlieric, heterogeneous gneisses, which in areas of poor exposure can be hard to distinguish from the Amîtsoq gneisses. The granodioritic gneisses contain inclusions of homogeneous hornblende + biotite dioritic gneiss. These granodioritic and dioritic gneisses resemble, and are continuous to the south-west with, the type Ikátoq gneisses of the Tre Brødre terrane.

In the Kapisigdlit area, the Ikátoq gneisses contain abundant inclusions of garnet + biotite ± sillimanite ± cordierite ± staurolite metasediment, banded amphibolite, metaquartzite, metagabbro, anorthosite and subordinate ultramafic rocks. The metasediments are locally found as inclusions in the gabbro and leucogabbro.

The metaquartzite apparently forms a single unit up to 50 m thick, which can be traced for many kilometres along strike, despite being intruded and disrupted by

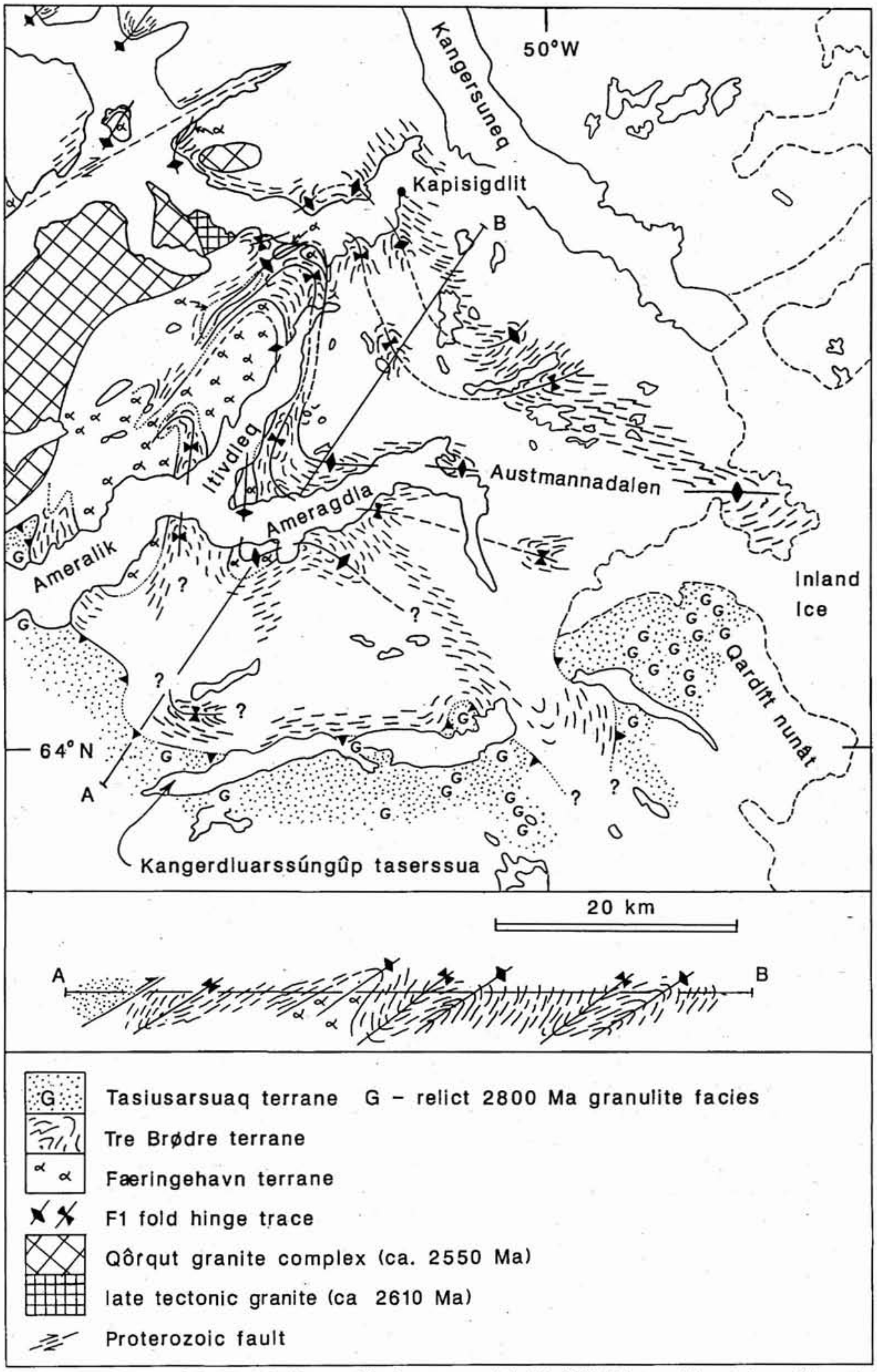

Fig. 1. Geological sketch map of the Kapisigdlit area. Blank areas not visited in 1987. Western 5 km of Ameralik on the figure was mapped by V. R. McGregor.

Ikátoq gneisses. The quartzite is commonly white, and consists of quartz-rich layers a few millimetres to a few centimetres thick, separated by thin seams containing mica and opaque minerals. Locally the quartzite contains finely disseminated chromian muscovite, giving it a green colouration. Two new localities were found (64° 15′ 40″N, 50° 5′ 15″W and 64° 23′ 45″N, 50° 1′ 5″W) where this colouration is sufficiently strong for the quartzite to be worked for 'greenlandite'. Previously this quartzite had been reported at only two localities (64° 20′ 20″N, 50° 10′ 50″W) in the Kapisigdlit area (found by V. R. McGregor), and on the island of Simiutâ, outer Ameralik (Nutman & Bridgwater, 1983). Detrital zircons separated from quartzite at these localities are early Archaean in age, showing that they are derived at least in part from erosion of ancient crustal rocks, such as the Amîtsoq gneisses (P. Kinny, personal communication, 1987; Schiøtte *et al.*, in press).

The gabbro and leucogabbro commonly show layering. This is interpreted as due to igneous structures transposed during later ductile deformation. The leucogabbros, gabbros and ultramafic rocks are interpreted as having been derived from a large stratiform intrusion at least 200 m thick, consisting of layered gabbro with interspersed ultramafic units at the base, followed upwards by leucogabbro with subordinate units of gabbro and anorthosite. No chromite-rich units were found in these rocks.

In the Kapisigdlit area, the Tre Brødre terrane does not contain inclusions of Amîtsoq gneisses, as seen locally in this terrane to the south-west (Friend *et al.*, 1987).

Tasiusarsuaq terrane. The south of the area consists of polyphase, nebulitic tonalitic gneisses containing scattered inclusions of basic rocks. These rocks are continuous with the Tasiusarsuaq terrane to the south and west, affected by granulite facies metamorphism at *c.* 2800 Ma (McGregor *et al.*, 1986; Friend *et al.*, 1987). In the Kapisigdlit area these nebulitic gneisses with their basic inclusions have brown-weathered patches up to 500 m across, with relict granulite facies assemblages. The gneisses with their basic inclusions adjacent to these patches have amphibolite facies assemblages. However, these gneisses have blebby textures attributed to recrystallisation under granulite facies conditions (McGregor *et al.*, 1986). The blebby textures have subsequently been transposed to varying degrees during deformation as the rocks were retrogressed under amphibolite facies conditions. On the south side of Kangerdluarssúngûp taserssua (fig. 1), helicopter reconnaissance showed that granulite facies assemblages with subvertical foliation are commonly best preserved on hill tops above 1000 m. At lower altitudes on the same hills, these granulite facies gneisses are totally retrogressed under amphibolite facies conditions, with their subvertical foliation transposed into a new southerly dipping foliation.

Tectonometamorphic evolution

Terrane boundaries. The boundaries between terranes are marked by zones, generally about 10 m wide, but in places up to 50 m wide. These zones consist of finely layered quartzo-feldspathic and quartz-rich rocks and mica and amphibole rich schistose rocks. The schistose rocks commonly contain thin disrupted pegmatite veins and feldspar grains up to 1 cm across, which may represent porphyroclasts. The finely layered quartzo-feldspathic rocks and quartz-rich rocks have mylonitic to ultramylonitic textures with superimposed blastomylonitic textures. By analogy with studies in the Tre Brødre area (Friend *et al.*, 1987), these rocks are interpreted as mylonites developed during the juxtaposition (emplacement) of the three terranes.

Emplacement of terranes and folding. The cross-section (fig. 1) shows that the Færingehavn terrane is structurally lowest, and that the Tasiusarsuaq terrane, affected by granulite facies metamorphism at 2800 Ma, is structurally highest. This is the same order of stacking of terranes found in the Tre Brødre area (Friend *et al.*, 1987), and indicates that the emplacement of the terranes must have occurred after 2800 Ma (the age of the granulite facies) and must have involved thrusting of the Tasiusarsuaq terrane over the other two terranes. Further evidence for this comes from south of Kangerdluarssúngûp taserssua (fig. 1), where granulite facies assemblages are least retrogressed on the tops of hills.

After their tectonic juxtaposition, the Færingehavn and the Tre Brødre terranes were folded together in large, tight to isoclinal folds (F_1). These folds were mapped by tracing the tectonic boundary between the Færingehavn and the Tre Brødre terranes, by using as structural markers large inclusion trains of metasediment, gabbro and leucogabbro in the Tre Brødre terrane, and by making extensive use of the sense of vergence of F_1 parasitic folds. Traces of the hinges of major F_1 folds are indicated on fig. 1. The F_1 folds have associated with them a mineral lineation parallel to their hinges.

The F_1 folds seem to die away southwards towards the Tasiusarsuaq terrane, so that the boundary between the Tasiusarsuaq and the Tre Brødre terranes has a simpler outcrop pattern than the boundary between the Færingehavn and the Tre Brødre terranes (fig. 1). To ex-

Table 1. Geological evolution of the Kapisigdlit area

Events common to all three terranes	
3	Intrusion of the Qôrqut granite complex at *c.* 2550 Ma. Intrusion of late tectonic granite at *c.* 2610 Ma.
2	Intrusion of granitic sheets correlated with the *c.* 2690 Ma Nipinganeq granite sheets of middle Ameralik. Formation of F_2 folds and subvertical shear zones.
1	Tectonic juxtaposition of first the Færingehavn and Tre Brødre terranes, and then the Tasiusarsuaq terrane with F_1 folding in the Færingehavn and Tre Brødre terranes. Events 1 and 2 occurred between 2550 and 2750 Ma, and were accompanied by amphibolite facies metamorphism, with retrogression of granulite facies assemblages in the Tasiusarsuaq terrane.
Evolution of the Tasiusarsuaq terrane (upper structural unit)	
2	Granulite facies metamorphism at *c.* 2800 Ma.
1	Intrusion of the protoliths of voluminous, predominantly tonalitic gneisses into a sequence of volcanic rocks and related intrusions. Deformation and high grade metamorphism. 2800 to 3000 Ma?
Evolution of the Tre Brødre terrane (middle structural unit)	
2	Intrusion of the dioritic and granodioritic protoliths of the Ikátoq gneisses, with deformation and metamorphism, between 2750 and 2800 Ma.
1	Formation of supracrustal rocks and intrusion of a gabbro - leucogabbro complex into them.
Evolution of the Færingehavn terrane (lower structural unit)	
3	High grade metamorphism at *c.* 3600 Ma?
2	Intrusion of protoliths of polyphase tonalitic Amîtsoq gneiss and granite sheets. High grade metamorphism and deformation. Pre-3600 Ma.
1	Formation of the Akilia (supracrustal) association. 3800 to 3850 Ma.

plain this, we suggest that the F_1 folds developed at structural levels below the Tasiusarsuaq terrane which acted as a rigid block as it was thrust into position. Therefore, if the Tre Brødre and Færingehavn terranes had already been emplaced by the time the Tasiusarsuaq terrane was thrust into position, their outcrop pattern would be controlled by F_1 folds. It is concluded that the Færingehavn and Tre Brødre terranes were emplaced first, and that subsequently the Tasiusarsuaq terrane was thrust into position (Table 1).

After the emplacement of all three terranes, but prior to intrusion of the *c.* 2550 Ma (Baadsgaard, 1976; Moorbath *et al.,* 1981) post-tectonic Qôrqut granite complex (fig. 1), non-cylindrical folds (F_2) developed. These folds commonly have an *en echelon* basin and dome form, are closed to tight, trend NNE and their axial surfaces dip to the north-west. F_2 parasitic folds are generally close to open, with the same style as the larger structures. Overprinting of F_2 folds onto F_1 folds is responsible for the complex outcrop pattern of the boundary between the Færingehavn and Tre Brødre terranes (fig. 1). F_2 structures seem to become more open and with a greater wavelength eastwards towards the Inland Ice. Mineral lineations coaxial with F_1 folds are rotated around F_2 structures (fig. 2). In the tighter F_2 structures in the west of the area, F_1 lineations are commonly rotated into near parallelism with F_2 fold hinges, giving the impression that new linear fabrics formed during the F_2 event.

Metamorphism. Since the intrusion of the Ameralik dykes in the middle Archaean, the Færingehavn terrane has not been affected by granulite facies metamorphism. Likewise, the Tre Brødre terrane has never been

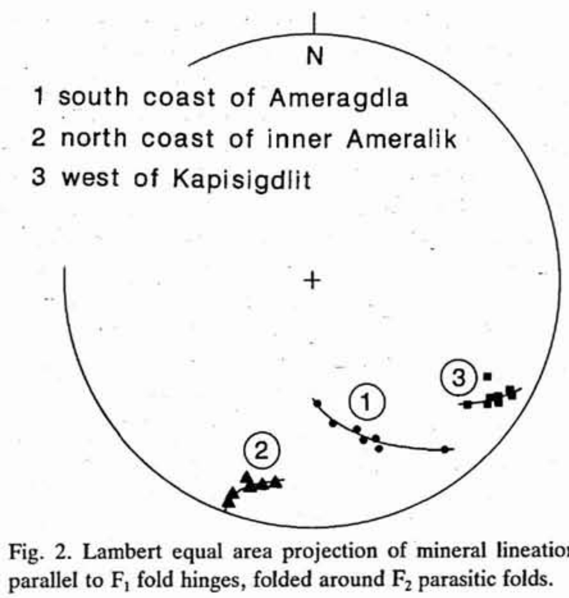

Fig. 2. Lambert equal area projection of mineral lineations parallel to F_1 fold hinges, folded around F_2 parasitic folds.

affected by granulite facies metamorphism. Amphiboles and sillimanite that are part of the amphibolite facies assemblages of the Færingehavn and Tre Brødre terranes generally have the mineral lineation parallel to the F_1 fold hinges. On the other hand, the Tasiusarsuaq terrane contains relict *c.* 2800 Ma granulite facies assemblages and amphibolite facies assemblages produced by retrogression of the granulite facies assemblages. Therefore most of the amphibolite facies assemblages in the area reflect *P, T* conditions during emplacement of the terranes and F_1 and F_2 folding later than 2800 Ma, but before 2550 Ma (Table 1). Simplistic models of crustal evolution such as those suggested by Wells (1979) for adjoining areas to the south, which involve a single post-tectonic metamorphic peak at *c.* 2800 Ma, cannot therefore be applied.

Late Archaean granites

Roberts (1979) described granitic sheets (Nipinganeq granites) from middle Ameralik, which give a Rb-Sr whole rock isochron age of *c.* 2690 Ma. These sheets are commonly composite, with pegmatitic margins and foliated gneissic cores. Many of them were intruded along active shear fractures. Similar sheets of granite were found in all three terranes to the east in the Kapisigdlit area. These granitic sheets post-date emplacement of the terranes and F_1 folding, but seem to be affected by F_2 folding. To the west of Kapisigdlit village there are swarms of pale, medium to fine grained granite sheets that are very weakly deformed. They seem to post-date F_2 folding. These sheets increase in abundance westwards, giving rise to the late tectonic granite (fig. 1), dated at *c.* 2610 Ma (C. R. L. Friend, unpublished data). Undeformed pegmatite sheets up to 10 m thick occur sporadically throughout the area, increasing in abundance westwards. They are correlated with the *c.* 2550 Ma Qôrqut granite complex (fig. 1). Dating of these granitic sheets provides further evidence that ductile deformation continued sporadically throughout the late Archaean with associated recrystallisation under amphibolite facies conditions, and did not cease at *c.* 2800 Ma.

Discussion

The investigation of the Kapisigdlit area supports the model of crustal evolution for the area to the south and west recently proposed by Friend *et al.* (1987), involving the separate evolution and then emplacement of three terranes in the late Archaean.

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Application of seismo-stratigraphic interpretation techniques to offshore West Greenland

James A. Chalmers

A pilot study is being conducted to determine if the use of seismo-stratigraphic interpretation techniques can increase the understanding of the geology of offshore West Greenland in order to reassess the prospectivity of the area.

During the period 1975 to 1979, a number of concessions offshore West Greenland were licensed to various consortia of oil companies to search for petroleum. Some 40 000 km of seismic data were acquired, all of which is now released. Five wells were drilled, all of them dry, and all concessions were relinquished by the industry by 1979.

The regional geology of offshore West Greenland has been summarised by Manderscheid (1980) and Henderson *et al.* (1981). They show the West Greenland Basin to consist of fairly uniformly westward dipping sediments bordered near the shelf break by a basement ridge. These authors used what may be termed 'conventional' techniques of seismic interpretation. However, since that time the techniques of seismo-stratigraphy (Vail *et al.*, 1977; Hubbard *et al.*, 1985) have become established. They are now being applied to study seismic data acquired during the mid-1970s.

Interpretation

Concession 26, held by a group operated by Amoco Greenland, lies south-west of Færingehavn on GGU 1:100 000 base map 63 S 51 30 (fig. 1). It was relinquished during 1978 without drilling. A seismic grid of roughly 3 × 5 km exists over the former concession. The data were acquired in 1975 and 1977 and are either 24- or 48-fold. None of the lines have been migrated. The

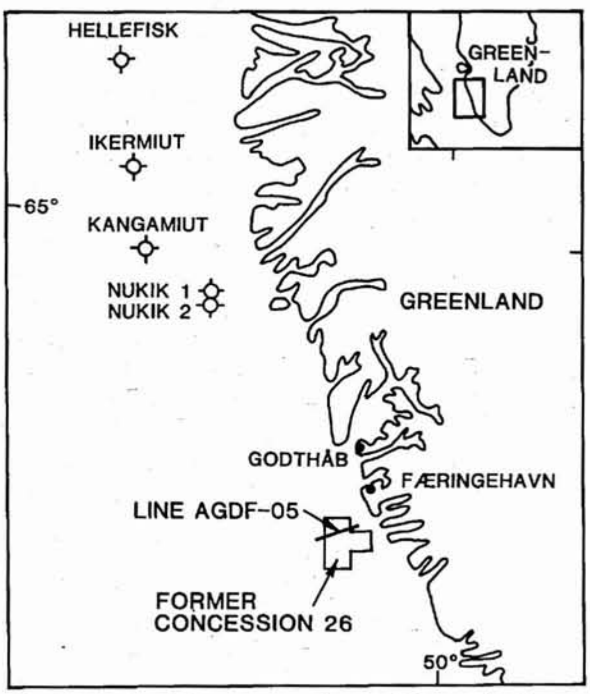

Fig. 1. Location of former concession 26 offshore West Greenland.

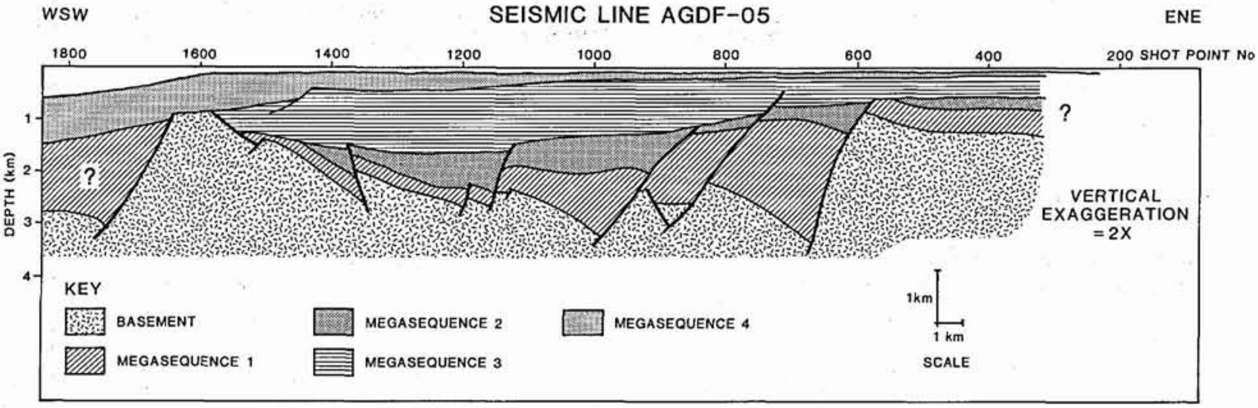

Fig. 2. Depth converted interpretation of seismic line AGDF–05 showing Megasequences 1 to 4.

area is a self-contained study area covered by a grid of fair quality seismic data and in which the most recent sediments are thin (see later). This allows easier penetration of seismic waves to deeper levels, which can perhaps be seen more clearly here than farther north.

Using the techniques described by Vail *et al.* (1977), depositional sequence boundaries were identified and traced round the grid. It was found that there existed four megasequences in the sense defined by Hubbard *et al.* (1985). These are shown in fig. 2 on a depth-converted interpretation of seismic line AGDF-05.

Megasequence 1 is a syn-rift sequence of unknown age. It lies within a series of coalescing half-graben. Within it a number of individual sequences can tentatively be identified, though it is unlikely that the depositional environment can be identified within these, due to the comparatively poor quality of the seismic data. Reprocessing of the seismic data could change this situation, however.

Megasequence 2 is also a syn-rift sequence. It overlies Megasequence 1, but the two rift phases have not been bounded by the same faults, and extension has been in different directions in places. This has resulted in space problems which have been resolved by the formation of thrust faults, as at shot point 1350 of fig. 2. Megasequence 2 is of uniform seismic character and seems to consist of only one sequence.

Megasequence 3 is a post-rift sequence which seems to correspond to the thermal phase (in the sense used by McKenzie, 1978) of the rift containing Megasequence 2. It contains four sequences within which it is probably possible to identify the depositional environments.

Megasequence 4 is a second post-rift sequence which overlies Megasequence 3 and truncates it with marked unconformity to the north and west. It again contains several individual sequences.

Ages of the sediments

There is no direct way to date these sequences at present. It may be possible to follow them northwards into the wells that have been drilled there, but that will require much more work than is at present underway. However, it is tempting to speculate that Megasequence 4 lies above the unconformity of Oligocene age which is widely identified around the Atlantic (Vail *et al.*, 1977). If so, then Megasequences 3 and 4 could be the rift and post-rift sequences associated with the Late Cretaceous onset of sea-floor spreading in the Labrador Sea (Srivastava *et al.*, 1981). This would mean that Megasequence 1 is of unknown earlier age. One can speculate as to how old, but it seems that no rocks of that megasequence have yet been sampled on the West Greenland Continental Shelf.

An alternative interpretation of the unconformity sequences could be that the unconformity between Megasequences 3 and 4 was produced by the change in tectonic regime at the start of sea-floor spreading in the Norwegian Sea. This would make it early Eocene in age (Talwani & Eldholm, 1977). If so, Megasequences 2 and 3 would still be associated with the onset of spreading in the Labrador Sea, at anomaly 32 time (Srivastava *et al.*, 1981). This would make them Late Cretaceous to Palaeocene in age. Yet again, however, this means that Megasequence 1 is of unknown age and provenance.

Work is still continuing on mapping these sequences and attempting to identify the depositional environments. A full report is planned for publication in 1988.

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Detailed investigation of the niobium-tantalum distribution within the Motzfeldt Centre, South Greenland

Bjørn Thomassen, Tapani Tukiainen and Karsten Secher

A one year programme for detailed investigation of selected parts of the Nb-Ta-U-bearing pyrochlore mineralisation within the Motzfeldt Centre, South Greenland, was initiated in June 1987. The aim of the programme is to delineate areas with a potential for high-grade accumulation of Ta-enriched pyrochlore which is important if international exploration and mining activity is to be attracted. It is also important for official decisions in relation to any application for concessionary activity. The programme is carried out as a joint project between GGU and Nunaoil A/S under supervision of GGU, and a major part of the funding is granted by the Mineral Resources Administration for Greenland.

On the basis of information obtained during survey activity from 1980 to 1986 (Tukiainen, 1986), six mineralised areas covering *c.* 1.5 km^2 were selected for detailed investigation by gamma-spectrometry and rock sampling (fig. 1). Most of the areas are situated on very steep slopes in alpine terrain with altitudes up to 1900 m a.s.l. Therefore, the gamma-spectrometric measurements were made by helicopter, and the sampling was carried out by mountaineers.

In total 20 people were engaged during the field season from the end of June to the end of August (Thomassen & Tukiainen, 1987). The authors functioned as project geologist, consultant and leader, respectively. The GGU station in Narsarsuaq was operated as the field base by J. Lau. The helicopter service was by Greenlandair Charter A/S, the geophysical survey by Global Earth Sciences Ltd. (England), and the mountaineers were provided by Garaventa AG (Switzerland).

Geology and mineralisation

The Motzfeldt Centre (1310 ± 10 Ma, Blaxland *et al.*, 1978) is one of the major central complexes in the Gardar Province of alkaline igneous activity. The centre belongs to the Igaliko nepheline syenite complex of which a general account was given by Emeleus & Harry (1970). Following the discovery of radioactive mineralisation in 1979 (Armour-Brown *et al.*, 1980), the centre has been the objective of intensive research. The detailed geological and radiometric mapping of the Motzfeldt Centre was initiated in 1982 as a part of the 'Syduran project' (Armour-Brown *et al.*, 1983), and gradually its potential for large-scale Nb-Ta mineralisation was recognised (Tukiainen *et al.*, 1984). The research was continued during the project 'Pyrochlore in alkaline intrusions of Greenland' which was partly funded by the European Economic Communities (Bradshaw, 1985; Tukiainen, 1986, in press).

The Motzfeldt Centre covers an area of *c.* 300 km^2. It is made up of multiple intrusions of syenites emplaced into the Proterozoic Julianehåb Granite and the overlying Gardar supracrustal rocks. The main igneous

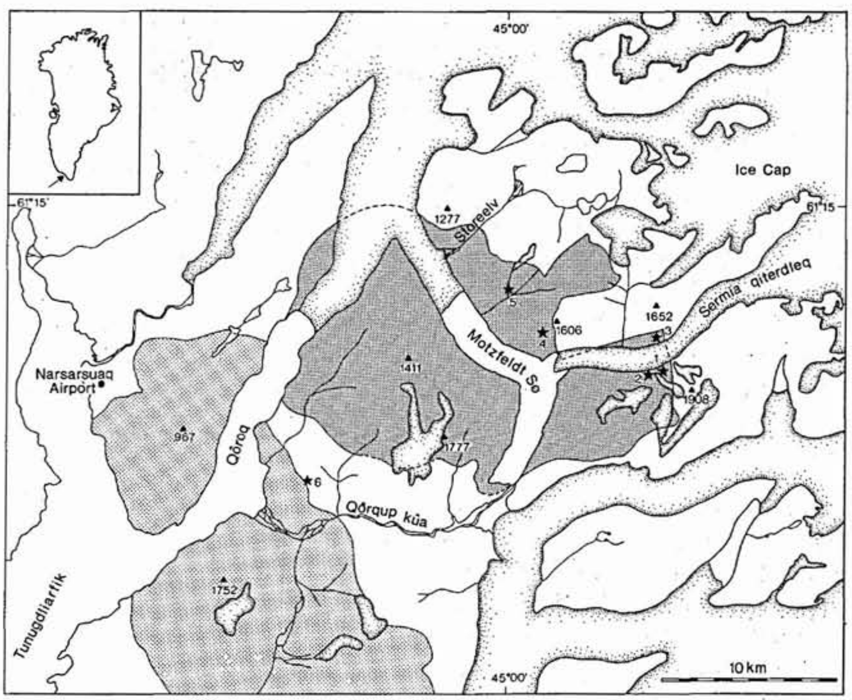

Fig. 1. Map of the Motzfeldt Sø area. The Motzfeldt Centre is dark grey, other Gardar intrusions are light grey. The investigated localities are indicated with stars.

phase (the Motzfeldt Ring Series) consists of a number of largely concentric, steep-sided, outward dipping units of predominantly peralkaline syenite and nepheline syenite which young inwards. It is the outermost of these units, the Motzfeldt Sø Formation, which hosts the mineralisation. The apparent intrusion mechanism was a combination of ring fracture and block subsidence. Post-dating the syenite intrusions the centre was dissected by a number of faults with displacements up to 6 km horizontally and 600 m vertically.

Large quantities of roofing sandstones and volcanics have been incorporated into the Motzfeldt Sø Formation. The volcanics are preserved as large rafts, but the sandstones have largely been assimilated and caused an outer zone of the Motzfeldt Sø Formation with a silica saturated composition. The formation underwent an extreme *in situ* differentiation, probably due to effective crystal fractionation, which resulted in the formation of a peralkaline residuum rich in volatile and incompatible elements. The peralkaline residuum gave rise to a complex of late peralkaline sheets of microsyenite and pegmatite, and to hydrothermal alteration with associated Th-U-Nb-Ta-Zr-REE mineralisation which increases in intensity outwards, especially towards the roof of the intrusion.

The mineralisation is probably the result of a combination of an incompatible element/volatile enriched magmatic residuum and an influx of silica and meteoric water, which resulted in a dramatic increase in oxygen fugacity, acidity, and hydrothermal activity. A volatile-saturated outer shell developed which facilitated the migration, accumulation, and precipitation of the incompatible elements.

The Motzfeldt Sø Formation hosts zones of economically interesting pyrochlore enrichment in altered syenite and in peralkaline microsyenite. The pyrochlore is enriched in Ta, U and LREE. The Ta content and the Nb/Ta ratio of the pyrochlore vary from 1.5 to 10.0% Ta_2O_5 and from 8 to more than 50, respectively, but in general Ta contents are higher in pyrochlore from altered syenite than in pyrochlore from peralkaline microsyenite. The pyrochlore shows also a marked compositional variation depending on its relative depth in the Motzfeldt Sø Formation. At the deeper levels it is enriched in Ta and Ca, whereas at the higher levels of the igneous column it is more enriched in Nb, U and LREE.

Airborne gamma-spectrometric survey

A gamma-spectrometric survey of the whole Motzfeldt Sø area in 1982 (Tukiainen *et al.*, 1984) and subsequent research (Tukiainen, 1986) has demonstrated a good correlation between radiometric anomalies with high U content and high U/Th ratio, and pyrochlore mineralisation. In order to outline the anomalies in detail and to facilitate the extrapolation of element concentrations beyond the sampled areas, the six most anomalous localities were reflown at the beginning of the season with a more advanced airborne survey system installed in a larger helicopter (fig. 1).

Equipment. The survey system with a total weight of 201 kg was installed in an Écureuil AS 350 B1 helicopter and consisted of a 256 channel gamma-ray spectrometer with a crystal detector of 16.6 litres of Tl-activated NaI crystals, a data acquisition system with tape deck, a six-channel analogue recorder, radar and barometric altimeters, and a video flight path tracking system. The various items of the survey system were fully synchronized.

Procedure. The survey was carried out by contour flying with a contour spacing of 30 m. The system was flown at a speed of 25–40 knots (45–72 km/h), and an average terrain clearance of 30 m was attempted. To improve the reliability of the radar altimeter recordings in areas with steep mountain slopes, the altimeter was tilted *c.* 30° against the mountain face to be surveyed. The fixed position and tilt of the radar altimeter on the pilot's side of the aircraft meant that the survey had to be carried out unidirectionally. In order to improve the spatial resolution of the survey, the counting time was set to 0.5 sec., which was considered adequate for the known high level of radiation and the large detector volume. The airborne system recorded the accumulated counts (total counts, K, Th and U windows) on magnetic tape together with radar-measured ground clearance, barometric altitude, real time and fiducial number, and generated an analogue chart of these data. The flight paths were recorded by a video camera which was installed below the aircraft. These recordings were used for the final flight path recovery. Navigation was based on contoured orthoscopic aerial photographs at a scale of 1:5000 supplemented by oblique photographs.

The energy calibration of the system was performed with a ^{137}Cs gamma-ray source. The background radiation level and instrumental drift were checked at the beginning and end of each flight. The background radiation test was carried out over sea water, and checks for instrumental drift and determination of the attenuation coefficients were made over a 2.5 km long test strip. Checks for cosmic radiation were made at the beginning and end of the survey programme.

Results. The survey involved 23 hours 52 minutes of flying time, of which 19 hours 25 minutes were used for the survey proper and the remaining time for attenuation and cosmic tests. A total of 454 line km were flown, and 63,443 measurements with an average point separation of *c.* 7 m were recorded over an area of *c.* 16.3 km^2 (Table 1).

A preliminary evaluation and interpretation of the data based on uncorrected raw data from the analogue charts was made immediately after the survey was completed. The anomalies were manually plotted on the flight track maps and used to guide the chip sampling programme.

The final processing by Global Earth Sciences Ltd. included correction of the spectrometric data for background variations, spectral interferences and ground clearance, and merging of this data with navigation information.

Detailed chip sampling

To obtain reliable data from the best mineralised areas in the Motzfeldt Centre, chip samples were collected systematically at five localities (1–5 on fig.1). At the first four localities the mountains are so steep (slope angles 50°–90°) that the sampling was performed by six mountaineers under supervision of the project geologist. The group was lodged in a mobile field camp consisting of three caravans. Selection of the localities was based on previous work (Tukiainen, 1986), with the main emphasis on tantalum mineralisation in altered syenite of the Motzfeldt Sø Formation.

Procedure. The mineralised localities were sampled as far as possible on a grid with the grid lines perpendicular to the contours. The spacing between grid lines was originally 50 m and the vertical distance between sample points in the lines was also 50 m. After the aeroradiometric survey had localized the anomalies, additional sampling on a 25 × 25 m grid was performed over the best anomaly at locality 1 (fig. 2), and a 25 × 25 m grid was used from the beginning at localities 3 to 5.

Table 1. Summary of results of aeroradiometric survey and chip sampling in 1987, Motzfeldt Centre, South Greenland

	Aeroradiometric survey		Chip sampling programme	
Loc.	Number of measurements	km² covered	Number of samples	km² covered
1	15,017	3.4	329	0.26
2			119	0.15
3	8,922	1.8	286	0.12
4	10,501	3.0	145	0.07
5	15,257	3.3	49	0.04
6	13,736	4.8	0	0.00
Total	63,443	16.3	928	0.64

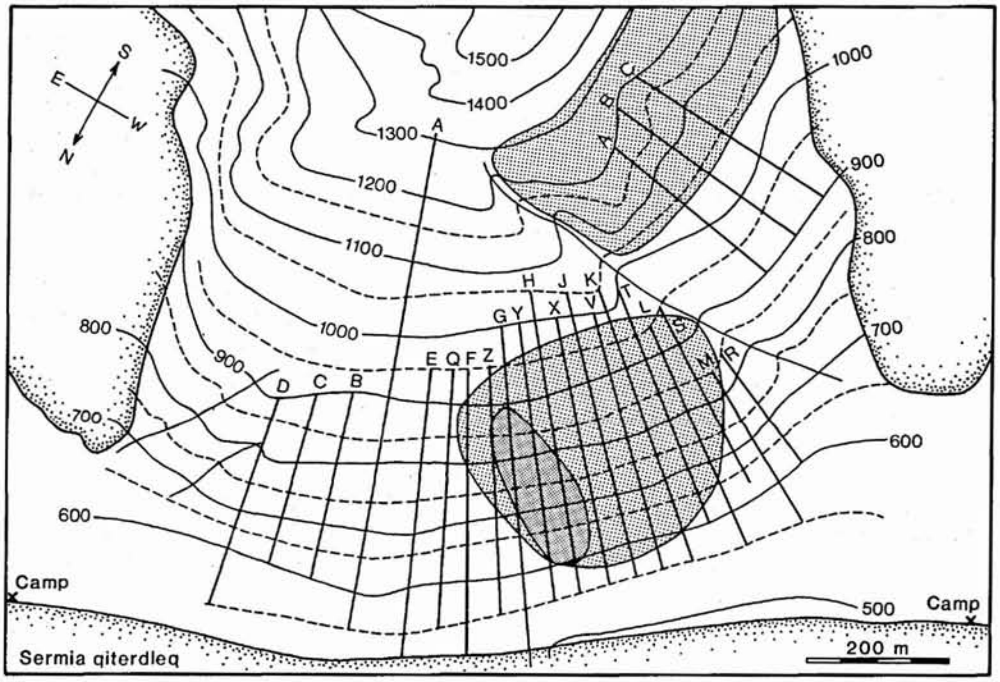

Fig. 2. Map of Qiterdleq South (locality 1). Aeroradiometric anomalies are hatched, and chip sampled grid lines are marked with letters.

The daily sampling started at the top of the anomaly, where the climbers went by helicopter or on foot from the field camp. They measured out the distances between the grid lines and descended over the mountainside in groups of two or three, as far as possible in straight lines. At each sample point (determined by a barometric altimeter) a 2 × 2 m cross was spray-painted together with a preliminary number. A chip sample and two small reference samples were collected inside the 4 m^2 defined by the cross, and thereafter a scintillometer measurement was performed. Working its way down in this manner, a two man group could typically collect 10 samples during a 10–12 hour working day.

The sample sites were plotted on enlarged oblique photographs by the project geologist from observation points in the middle of the glacier. With a × 16 binocular mounted on a tripod it was possible to locate most sample points, and also to direct the sampling via a portable VHF radio.

In a laboratory established in Narsarsuaq, the chip samples were run through a jaw crusher and a mill, after which 100 g large splits were shipped directly to two commercial chemical laboratories in Canada. The samples are to be assayed for Nb, Ta, Th and U, and analysed for Be, Ce, La, Li, Mo, Sn, Y and Zr.

Results. Most of the radiometric anomalies defined in localities 1 to 5 by the aeroradiometric survey were covered by the chip sampling programme. An area with overhanging rock at locality 3 could not be sampled, and at other places a combination of steepness and loose rocks made sampling too dangerous, but on the whole, a satisfactory coverage was obtained at the main targets, localities 1 to 3. At localities 4 and 5, where several anomalies occur, the best anomaly was selected for detailed sampling. Locality 6 was not visited.

A total of 928 chip samples were collected over an anomalous area of 0.64 km^2 (Table 1). The sample weights ranged from 1.0 kg to 2.5 kg with an average of 1.8 kg.

Additional sampling

Bulk sample. With a view to possible beneficiation tests, a 200 kg sample was collected by means of drilling and blasting at locality 1. It consists of relatively fresh aplitic microsyenite or syenite with a high radiation.

Alluvial sands. In order to evaluate the Nb-Ta-mineral placer potential, 10 litre sediment samples were collected from the main drainage systems and the plains bordering Motzfeldt Sø. Heavy mineral concentrates were panned out from 29 stream-sediment samples, while four samples from a large glaciofluvial plain will be analysed untreated.

Concluding remarks

The aeroradiometric survey successfully reduced the areal extent of the previously known gamma anomalies by about 50%. This was due to the use of equipment with better resolution and a helicopter with better performance, compared with the 1982 survey. Consequently, smaller areas had to be covered by the systematic chip sampling, which enabled the collection of larger samples in closer spaced grids than originally planned.

The acquired geophysical and geochemical data will be treated statistically in co-operation with IMSOR (Institute of Mathematical Statistics and Operations Research), Technical University of Denmark, and will be interpreted in a final report.

Acknowledgements. The authors thank J. Lau and O. Plesner for invaluable support from the base camp at Narsarsuaq; E. Dam and I. Rytved for their spirited contribution in the field laboratory; L. Thorning for assistance during the geophysical survey; J. L. Pedersen and L. Melchior Larsen for comments on the report; and finally the Swiss mountaineering team and the British geophysical staff for professional and enjoyable co-operation.

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Greenland ice cap aeromagnetic survey 1987: completion of the survey over the southern end of the Greenland ice cap

L. Thorning, M. Bower, C. D. Hardwick and P. Hood

The Geological Survey of Greenland (GGU), the Geological Survey of Canada (GSC), and the National Aeronautical Establishment (NAE) of the National Research Council of Canada are cooperating in the GICAS project. The objective of the project is to achieve a regional, aeromagnetic coverage of the Greenland ice cap, and to produce magnetic anomaly maps for use in research on the large scale geological structures. Field work was carried out in 1983, 1984, and 1985 (Thorning *et al.* , 1984, 1985, 1986). The work reported in this note was originally planned for April 1986, but for technical reasons it had to be postponed until April 1987. Consequently, the subsequent processing of all the data into a regional magnetic anomaly map has been correspondingly delayed.

Field work

In 1985 parts of the southernmost ice cap were covered (Thorning *et al.*, 1986), and the 1987 operation was planned to make the coverage complete over this part of the ice cap. The operation was carried out as part of a larger programme to compile a magnetic anomaly map of North America, which also involved work in northern Canada and the Nares Strait region between Canada and Greenland, and a tie-line from the North Greenland continental shelf over the ice cap along its western margin to Søndre Strømfjord. Tie-lines from Alert to Svalbard to Iceland had to be abandoned due to unfavourable weather conditions around Svalbard.

The flights for the GICAS were based at Søndre

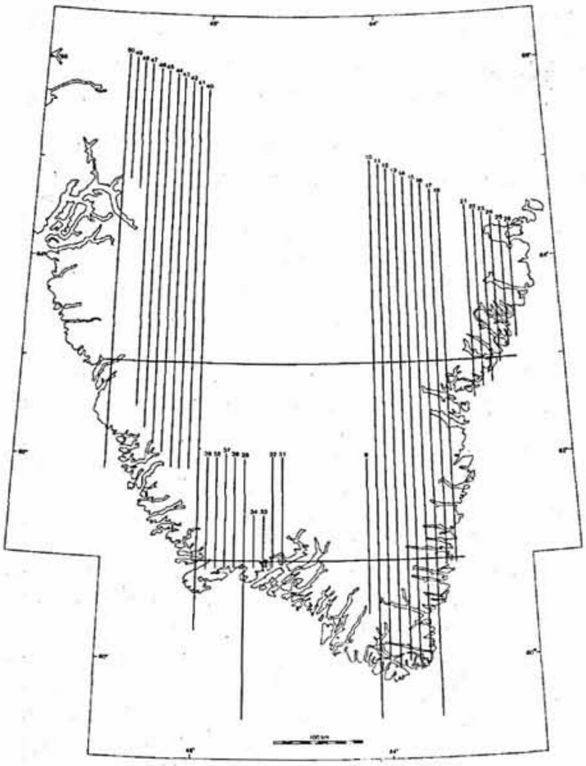

Fig. 1. Schematic representation of lines flown in April 1987. Approximately 8000 line km, line spacing approximately 10 km, flight altitude 1000 ft over ice surface or tops of mountains in ice-free, coastal areas.

Strømfjord airport, and Narsarsuaq was used for refuelling. A base magnetometer was operated in Søndre Strømfjord during flights to record diurnal variations.

The aeromagnetic data comprise both total field and three gradients (Hardwick, 1982). The NAE Convair 580 aircraft (C-FNRC, Research-9), already well equipped for navigation over the ice cap, carried a new global satellite navigation system.

The lines flown in April 1987 are shown in fig. 1. As in previous surveys the line separation is approximately 12 minutes of longitude which, at the latitude of Narsarsuaq, corresponds to 10.9 km and at the northern end of the area to 9.5 km. Approximately 8000 line kilometres were flown, including two additional east–west tie-lines. The coverage is now nearly complete over the southern part of the ice cap, with only a few gaps between the lines which it was not economic to fill out.

Further work

The data are undergoing post-flight processing at NAE, and will be transferred to GGU for further processing in the spring of 1988. In the field, inspection of the data confirms the observations from the 1985 survey (Thorning *et al.*, 1986) and closes the gaps in the coverage of major geological features. All data acquired in 1983–1987 (fig. 2) will be compiled into a magnetic anomaly map covering the entire area. No flying is contemplated in 1988, but it is hoped later to survey the triangular area north of Angmagssalik up to 71° north and across the ice cap.

Acknowledgements. Besides the authors the following participated in the field operations: W. T. Chevrier (aircraft captain), J. Aitken (co-captain), T. Slack, G. Hoftyzer, W. Budarick (project crew), M. Larsen (maintenance crew chief), S. O'Connor and D. Blythe (maintenance crew) all from NAE.

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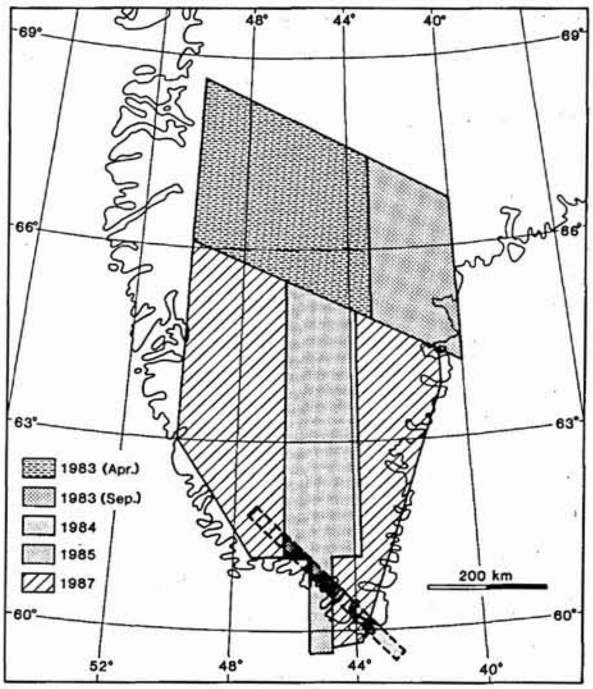

Fig. 2. Summary of Greenland ice cap aeromagnetic survey coverage 1983–1987.

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Reconnaissance investigations in the Skjoldungen region, South-East Greenland

Troels F. D. Nielsen and Jan C. Escher

From 1 July to 25 August 1987 a GGU expedition made reconnaissance investigations between 62°N and 64° 20′N in South-East Greenland. The programme was a continuation of the investigations in the Ammassalik/Angmagssalik district in 1986 (Kalsbeek & Nielsen, 1987) and the completion of the fieldwork describing areas in East Greenland between 62° 30′N and 65° 45′N for the planned map sheet (no. 14) in the 1:500 000 geological map series. The Skjoldungen district was known mainly from previous boat-supported work in the coastal areas as described by Bridgwater & Gormsen, 1969; Andrews *et al.*, 1971, 1973; Bridgwater *et al.*, 1976; Escher & Nielsen, 1982, 1983; Nielsen & Escher 1985 and Escher *et al.*, 1986.

Organization

The investigations were carried out as in 1986 in close co-operation with a team from the Geodetic Institute, Copenhagen, led by T. I. Hauge Andersson.

During an earlier airborne reconnaissance a camp site with a natural airstrip was located in the innermost part of Kagssortôq fjord (fig. 1). All equipment and personnel were transported to the base camp with a de Havilland Twin Otter (chartered by GI from Flugfélag Norðurlands hf., Iceland) from Kulusuk airfield near Ammassalik. The field work was supported by an Aerospaciale Ecureil (AS 350 B1) helicopter (chartered by GGU from Uni-Fly, Denmark). Both aircraft were chartered through Greenlandair Charter A/S and shared by GI and GGU.

The base camp was manned by P. Bay (materials, GGU), R. Fedder (cook, GGU), J. Wulff (radio communications, shared by GI and GGU) supported by A. Petersen (GGU) and P. B. Andersen (GI), both students. The helicopter was operated by two crews: S. Forsstedt (pilot) and S. Nielsen (maintenance) during the first three and the last two weeks of the season and by U. Stoller (pilot) and C. Jørgensen (maintenance) between 23 July and 12 August. The preparation for the field work and co-ordination in the field were undertaken by T. F. D. Nielsen (GGU) in co-operation with T. I. Hauge Andersson (GI).

The group of geologists included five two-man teams (fig. 1): a northern mapping team: J. C. Escher (GGU) and M. J. Ryan (Portsmouth Polytechnic); a southern mapping team: B. Chadwick (University of Exeter) and B. J. Walton (Portsmouth Polytechnic); two interchangeable mapping teams: M. T. Rosing (Geologisk Museum, Copenhagen), V. N. Vasudev (Department of Mines and Geology, Bangalore, India), T. F. D. Nielsen (GGU) and A. Petersen (GGU); and an age dating team through the last four weeks of the season: F. Kalsbeek (GGU) and P. N. Taylor (Oxford University). Reporting on mineral occurrences was assigned to V. N. Vasudev and M. J. Ryan.

The northern and southern reconnaissance teams were mainly based in field camps and were supported by reconnaissance with a helicopter and rubber dinghy. The rest of the group worked mainly from the base camp, also supported by both a helicopter and rubber dinghies.

Geology

The area investigated during the field season broadly corresponds with the Archaean terrain of South-East Greenland (Andrews *et al.*, 1973) which is bounded to

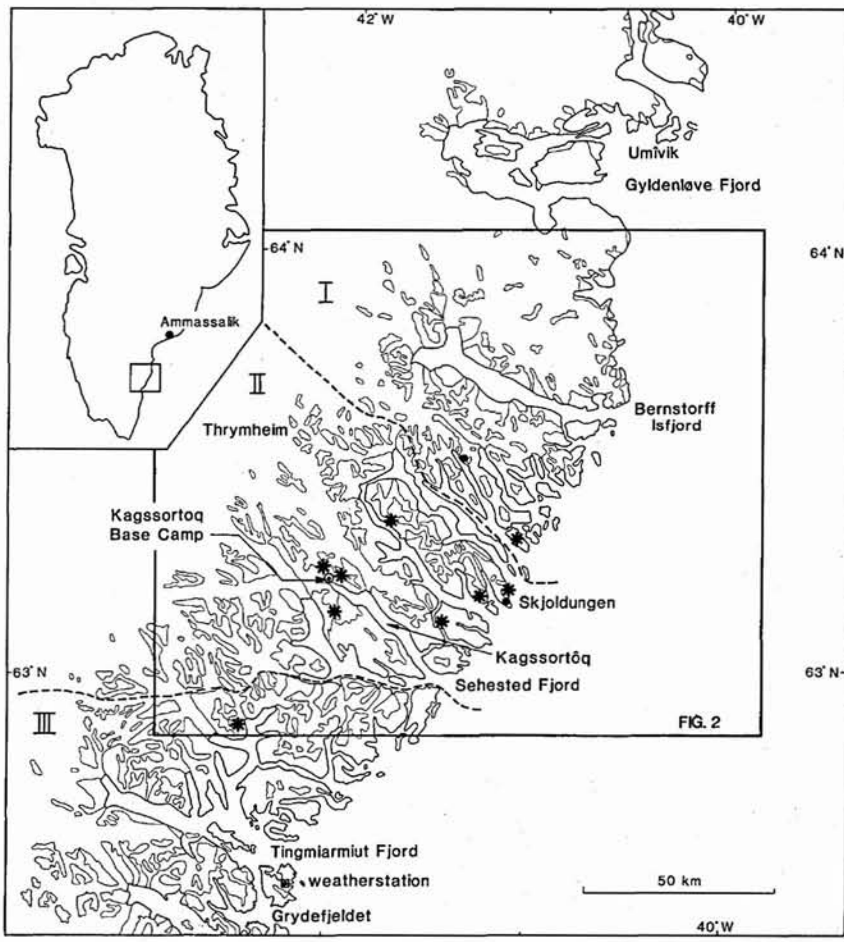

Fig. 1. The region investigated during field work 1987. The areas investigated by different teams are indicated: I J. C. Escher & M. J. Ryan; II V. N. Vasudev, M. T. Rosing & T. F. D. Nielsen; III B. Chadwick & B. J. Walton. Stars: Locality for sampling of material for radiometric age determination by F. Kalsbeek & P. N. Taylor. Filled circles: detailed investigations of supracrustal sequences.

the north at 64° 20′N by the Proterozoic 'Ammassalik mobile belt', previously called the Nagssugtoqidian of East Greenland (Bridgwater, 1976; Kalsbeek & Nielsen, 1987; Escher *et al.*, in press; Chadwick *et al.*, in press) and to the south at *c.* 62° 15′N by the acid volcanics and sediments of the Proterozoic Ketilidian mobile belt (Andrews *et al.*, 1971, 1973).

Reconnaissance age dating from a variety of rocks (Escher *et al.*, 1986) suggests that most of the gneisses were formed in the course of a major metamorphic-tectonic event 3000–2600 Ma ago. No evidence of early Archaean events has been observed.

A short description of the region, which has been divided into four areas each with its own geological characteristics, is given below. A detailed report of the summer's work will be given in a separate volume in GGU's report series.

Gneisses and supracrustal rocks

The region comprises gneissic and supracrustal rocks (fig. 2). The gneisses, which are the most common rock type, range from tonalitic to granitic and contain variable amounts of mafic inclusions. In the field they have been mapped as agmatitic gneisses and gneisses lacking inclusions, or with only a few. The bulk of the inclusions consist of amphibolites and ultramafic rocks and they have presumably been derived from supracrustal units. Together with the supracrustals they possibly represent the oldest rocks of the region. The supracrustals mainly comprise banded amphibolites with thin layers of ultramafic rocks, garnet-sillimanite paragneisses and quartzites. No igneous layering or pillow lava structures have been observed in the supracrustal sequences, but the impression is that most of the amphibolites are of volcanic origin.

Orthopyroxene has been identified in most of the gneisses and supracrustals throughout the region, even though most rocks have suffered retrogression to a variable extent.

The coastal area between Gyldenløve Fjord and Skjol-

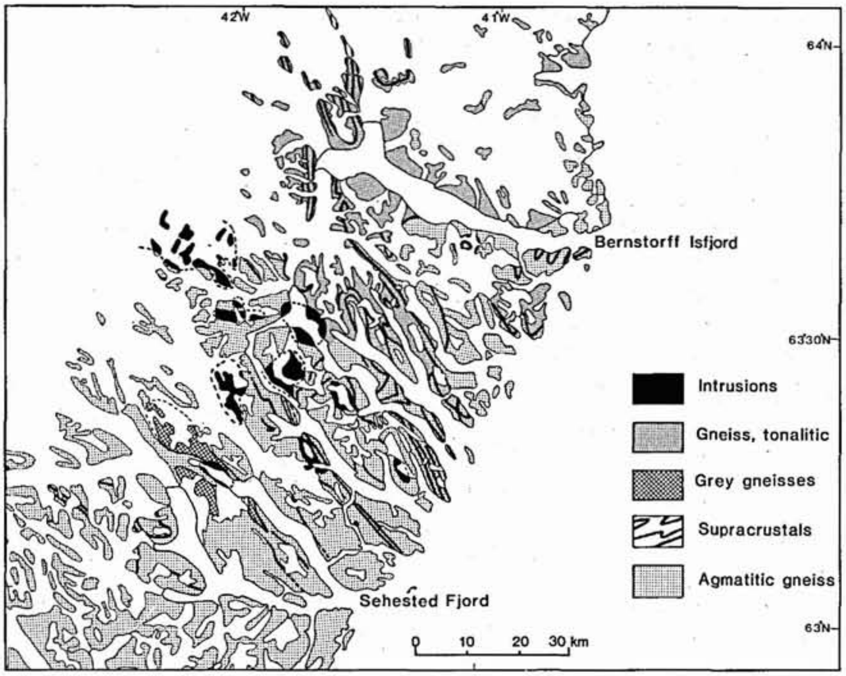

Fig. 2. Simplified geological map of the central part of the investigation area between Sehested Fjord and Bernstorff Isfjord showing the most important rock types.

dungen. Many supracrustal units can be traced along strike for 30 km or more, permitting large-scale interference fold patterns to be distinguished, which shows that the area has been affected by at least three major phases of deformation. The contact between gneisses and supracrustals is generally sharp and concordant; where local discordances occur, they are of tectonic origin. At a few localities a specific suite of lithologies occurs in the same order across strike: agmatitic gneiss, supracrustal amphibolite with thin ultramafics, metasediments and tonalitic to granitic gneiss with only few mafic inclusions. Folded and discordant trains of amphibolite inclusions show the occurrence of a few early dykes in the gneiss complex. The rocks are of amphibolite facies, but to the south near Skjoldungen orthopyroxene is frequently observed in the agmatitic gneisses and supracrustals.

The area along the margin of the Inland Ice between Gyldenløve Fjord and Skjoldungen. In contrast to the coastal area, the area along the margin of the Inland Ice has only a small proportion of agmatitic gneiss. The most common rock types are tonalitic-granodioritic gneisses and foliated granites, which all intrude banded amphibolites, ultramafic rocks and occasionally thin layers of metasediments. The supracrustals occur as elongated kilometre-sized rafts in the gneisses and granites. Regional folding has also here resulted in large-scale interference patterns, and a few large, recumbent, isoclinal fold hinges are well exposed in steep cliffs. The rocks are of amphibolite facies, but to the south they become brownish weathered and orthopyroxene bearing.

The Skjoldungen – Thrymheim area. The area consists of a *c.* 35 km wide zone that runs NW–SE from the Inland Ice to the sea and includes brownish-weathered agmatitic gneisses, grey gneisses, thick units of supracrustal rocks and a suite of intrusive complexes. The agmatitic gneisses, which form the main rock unit of this area, are rich in mafic inclusions. In parts of the area they have been partially remelted, and the mobilisates appear to intrude both agmatitic gneisses and supracrustals. At two localities in the paragneisses a fine layering of possible primary origin is preserved. As in the northern area, the supracrustal amphibolites and metasediments are locally bounded by agmatitic gneisses on the one side and tonalitic to granitic gneisses (poor in inclusions) on the other side. Some of the supracrustal belts can be followed up to 20 km along strike, but in the southern part of the area they are rapidly disrupted and form trains of xenoliths in the agmatitic gneiss. Locally the matrix is very rich in quartz and has presumably been derived from remobilised quartz-rich metasedimentary paragneisses.

Up to 10 km wide homogeneous sheets of syenitic to granitic gneisses occur in the agmatitic gneisses. These 'grey gneisses' lack, or contain only a few, mafic inclusions and may locally contain deformed, irregular, amphibolite dykes. Pegmatites related to these gneisses are often quartz-poor but are rich in large green amphibole.

The Skjoldungen area is characterised by a suite of intrusive bodies of gabbroic, dioritic, granitic and syenitic rock types. Similar intrusions have not been observed in the northern and southern parts of the region. They range from somewhat deformed to perfectly well preserved intrusions with well preserved primary magmatic textures. The largest gabbro and syenite complexes are found in the north-western part of Skjoldungen, whereas reddish-brown weathered granites dominate the nunatak area of Thrymheim. At present it is not clear whether the intrusions belong to several periods of plutonic activity or if they represent a suite of late- to post-tectonic intrusions.

The rocks in the coastal area from Skjoldungen and southwards to around Sehested Fjord are of granulite facies. In the rest of the area they are mainly amphibolite facies, but the presence of orthopyroxene indicates that at one time the agmatitic gneisses and the supracrustals attained granulite-facies conditions.

The region between Sehested Fjord and Tingmiarmiut weather station. This area is characterised by monotonous masses of agmatitic gneisses which represent about 95% of the rocks. The matrix is generally tonalitic and mafic inclusions are generally well rounded. An agmatised gabbro complex was also observed in the area. The only continuous supracrustal unit is found on Grydefjeldet west of Tingmiarmiut weather station. It is dominated by deformed amphibolites and minor garnet-sillimanite-bearing paragneisses. The belt can be followed over about 10 km and forms a major fold. Orthopyroxene has been identified in most parts of the area, but apart from the area around Sehested Fjord which is of granulite facies the rocks are of amphibolite facies. Many of the gneisses are epidote and K-feldspar bearing, as a result of the late intense shearing described below.

Alkaline complex

By a stroke of luck the base camp was erected within the perimeter of a previously unknown alkaline complex. Only a small part of the complex is exposed on the shores of Kagssortôq. It is about 8 km^2 in area and is composed of ultramafic, melteigitic, ijolitic and urtitic interlayered sheets cut by søvitic carbonatite and syenite dykes and pegmatites. The complex intrudes shear zones that affect all the gneiss types and most of the intrusions mentioned above. The age of the complex has not yet been determined.

Dykes and sheets

The earliest dykes are a few N–S orientated noritic dykes with characteristic interfingering contacts with the agmatitic gneisses. The dykes, which have not been identified north of Skjoldungen, appear fresh. They are followed by a dominant dyke generation of 20–80 m wide ENE–WSW orientated vertical dolerite dykes. South of Bernstorff Isfjord the dykes are unaltered magmatic rocks but northwards they become increasingly metamorphosed with amphibolitisation of their margins. In the Umîvik area at 64° 20′N the dykes are severely epidotised and amphibolitised and still further to the north in the Ammassalik Mobile Belt the same dykes are strongly deformed (Bridgwater & Gormsen, 1969).

Gently dipping green appinitic to carbonatitic sheets occur in the Tingmiarmiut region. They are of possible Ketilidian (middle-Proterozoic) age. In the same region a few NE–SW basaltic dykes of possible Gardar age (*c.* 1200 Ma) have been observed.

The youngest of the dyke generations consists of 10 to 40 m wide N–S trending dolerite dykes. They are often plagioclase phyric and very well preserved and are unaffected by shearing (see below). Similar dykes in the Umîvik area are regarded as Tertiary (Hall *et al.*, in press).

Shear and crush zones

A characteristic of the Archaean belt in South-East Greenland is the large number of shear and brittle crush zones. The zones affect gneisses and supracrustals and some of the intrusions in the Skjoldungen area. Large areas have been intensely affected, which hinders observation work during helicopter reconnaissance. Most of the zones are retrogressed to greenschist facies and are heavily mineralised by K-feldspar and epidote. A set of NE–SW trending zones cuts a NW–SE trending set. The alkaline complex is emplaced at a junction of major NW–SE and NE–SW shear zones, which it post-dates.

Mineral occurrences

Only a few mineral indications of possible economic interest have been noticed during the reconnaissance, and these are related to the supracrustal rocks. Some of the paragneisses are very rich in sillimanite (up to 50%) and some of the quartzites contain only a little sillimanite. Many of the paragneisses and the amphibolites contain sulphides, mostly pyrite and pyrrhotite. The alkaline complex contains small amounts of søvite carbonatite and has no anomalous radioactivity. Some of the søvites are rich in large apatite crystals and some of the pegmatitic feldspars show weak schillerization and contain massive sky blue cancrinite.

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Sedimentological studies of the fluviatile-shallow marine Upper Triassic to Lower Jurassic succession in Jameson Land, East Greenland

Gregers Dam

A three-year research fellowship programme supported by British Petroleum Development, London, was initiated in the summer of 1987. The main subject of the study is the Upper Triassic to Lower Jurassic succession in Jameson Land, East Greenland. This stratigraphic interval includes the Kap Stewart and the Neill Klinter Formations which have many features in common with some of the largest coeval hydrocarbon reservoir formations known in N.W. Europe (e.g. Statfjord field).

The core of the project is a lithofacies analysis but ichnology, palynology, source-rock analyses and porosity/permeability analyses will be included where relevant. If possible, corresponding intervals from the Norwegian continental shelf will be included in the project. The aim of the project is:

(1) to provide detailed and regional facies models for the two formations. Special stress will be laid on the physical stratigraphic relations in order to ascertain if regional unconformities are present and what order of magnitude they may represent.

(2) to establish a reservoir geological model which might help in the understanding of comparable reservoirs in the Norwegian-Greenland region. Particular attention will be paid to the geometry of individual sand bodies.

Field work

The field work in 1987 was carried out by one geological team and lasted for 5 weeks from late July to late August. Four localities were visited along the eastern margin of Jameson Land and in the southern part of Scoresby Land (fig. 1). Sedimentological studies involved detailed lithofacies and ichnofacies analyses as well as sampling for palynofacies analyses and porosity and permeability measurements.

Regional setting

The Mesozoic basins of East Greenland were formed in a failed rift system that was initiated in Carboniferous and Permian times (Surlyk *et al.,* 1981). The rift is orientated north–south and is cut by major NW–SE cross-faults that developed in the Jurassic and led to progressive northward-stepping downfaulting of blocks (Surlyk, 1977). Jameson Land is situated on the block farthest to the south in the exposed part of the graben system. This is the largest block and has the most complete Mesozoic succession which is at least 5 km thick (Surlyk *et al.,* 1981).

Throughout the Triassic the block was characterised mainly by continental rift deposition in an arid climate (Clemmensen, 1980). In Rhaetian – Hettangian times the basin was affected by cross-fault activity. To the north the basin was bounded by a NW–SE cross-fault in Kong Oscars Fjord (fig. 1). The southern boundary is unknown, but the continental fluvial deposits show that the palaeodrainage was towards the north, suggesting that a cross-fault was operative in Scoresby Sund (Surlyk *et al.,* 1981). The basin was bounded to the west by the same major fault zone as the Triassic basin and to the east probably by a N–S elongated landmass including the present day Liverpool Land (Surlyk *et al.,* 1981). The Rhaetian – Hettangian phase was dominated by fluvial and restricted marine deltaic sedimentation of the Kap Stewart Formation (Sykes 1974a; Clemmensen, 1976). The flora and presence of thin coal seams and rootlet horizons suggest that the climate had now become temperate and humid (Pedersen, 1976; Surlyk *et al.,* 1981).

During Pliensbachian – Toarcian times the basin was transgressed by tidal and shelf deposits of the Neill Klinter Formation (Sykes, 1974b). These deposits are also restricted to Jameson Land and Scoresby Land and indicate that the basin was bounded by the same major fault zones as during Rhaetian – Hettangian times. However, the new data suggest that there may have been an opening of the basin to the south.

During Jurassic times the depositional environment changed from the tidal deposits of the Neill Klinter Formation to a shallow siliciclastic shelf and finally to a wide muddy shelf (Surlyk *et al.,* 1981).

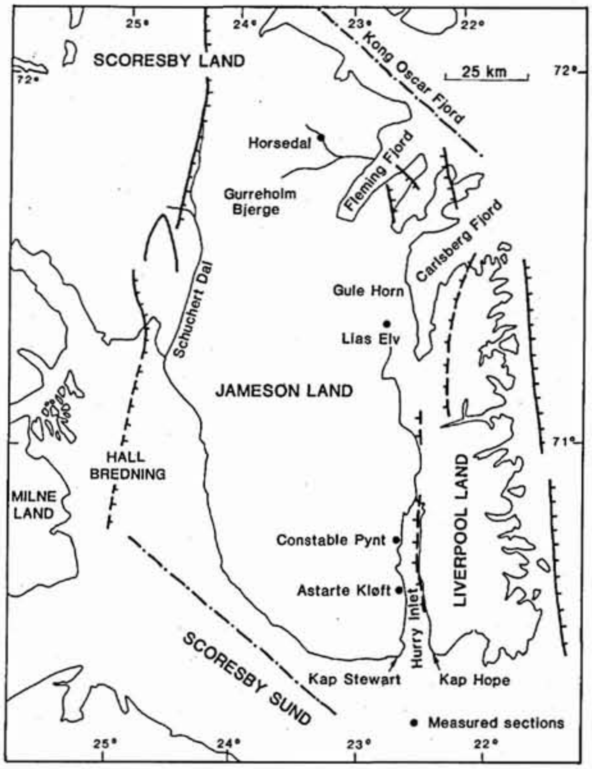

Fig. 1. Map of Jameson Land showing the localities of measured sections and major faults based on Surlyk *et al.* (1981).

Kap Stewart Formation

The Rhaetian – Hettangian sediments of the Kap Stewart Formation (Rosenkrantz, 1929) were studied in two areas along the west coast of Hurry Inlet (Astarte Kløft and Constable Pynt) and in Horsedal in southern Scoresby Land (fig. 1). The sediments along Hurry Inlet are known for their well-preserved and extensive flora described in a series of papers by T. M. Harris, published from 1926 to 1937. Harris (1937) divided the upper 'plant-bearing' series of the formation into two macroplant zones; a lower *Lepidopteris* Zone and an upper *Thaumatopteris* Zone. Pedersen & Lund (1980) made detailed palynological studies on the same localities in order to obtain a firmer dating of the macroplant zones of Harris by comparing spores and pollen from these zones with well-dated microfloras from northwestern Europe.

The formation is approximately 200 m thick along Hurry Inlet but well-exposed sections are present only in the upper 100 m in Astarte Kløft. At Constable Pynt the succession is poorly exposed but a virtually complete section through the formation is displayed. In Astarte Kløft and at Constable Pynt, fluvial-dominated delta plain facies associations occur which comprise fluvial-distributary channel sandstone, bank sandstone and interdistributary bay siltstone. Sykes (1974a) interpreted the sediments as being deposited in a low sinuosity non-braided river environment. The bank deposits occasionally show rootlet horizons; and plant debris and thin coal seams often occur in the interdistributary bay sediments. Palaeocurrent data from the distributary channels, although often very complex, suggest a general eastern to southern source area, the channels probably draining a landmass over present day Liverpool Land and a more southern area during this period.

In southern Scoresby Land the Kap Stewart Formation has an estimated thickness of approximately 350 m (Surlyk *et al.*, 1973). In Horsedal the uppermost 265 m are accessible often with good exposures. The sediments are characterised by wave-dominated delta-facies associations and have been interpreted by Clemmensen (1976) as tidally influenced deltaic deposits. Delta plain facies comprise fluvial-distributary channel sandstone and interdistributary bay mudstone and siltstone. Very little plant debris has been found in Horsedal. However, rootlet horizons and thin coal seams occur locally. The delta-front facies comprises 4–8 m thick coarsening-upwards sequences. The fine members at the base of the delta front are dominated by siltstone often with storm sand layers. The fine members grade up into well-sorted fine- to medium-grained sandstone which exhibits wave ripple lamination, horizontal stratification and cross-bedding. The sequences are often erosively overlain by medium- to coarse-grained fluvial-distributary channel sandstone. Sequences not topped by fluvial-distributary channels may represent longshore shoals formed under the influence of wave action. Palaeocurrent data from distributary channels suggest a source area to the north-north-west. Crest-line orientations of wave ripples indicate that the palaeo-coastline had a general E–W to NW–SE trend.

No invertebrate body fossils were found in the formation, and only a few unidentifiable trace fossils were observed at the three localities examined. The upper boundary to the overlying Neill Klinter Formation is well exposed in Horsedal and seems to be much more gradual here than along the Hurry Inlet where the Kap Stewart Formation is sharply overlain by probable shoreface deposits of the Rævekløft Member.

Neill Klinter Formation

The Pliensbachian – Toarcian sediments of the Neill Klinter Formation (Rosenkrantz, 1929; Lund & Pedersen, 1985) were studied in three areas; at Neill Klinter (Astarte Kløft and Constable Pynt), in Lias Elv (east of Carlsberg Fjord), and in southern Scoresby Land (fig. 1). The formation consists of three members (Surlyk *et al.*, 1973): the Rævekløft Member (base), the Gule Horn Member and the Ostrea Elv Member (top). These three members will be considered in turn.

Rævekløft Member. Deposition was initiated with the probable shoreface deposits of the Rævekløft Member. This member is approximately 9 m thick at Constable Pynt, thinning northwards to disappear entirely south of Carlsberg Fjord (Surlyk *et al.*, 1973). It is composed of poorly-sorted medium- to very coarse-grained sandstone and contains a rich Pliensbachian marine fauna, particularly in the lower part (Rosenkrantz, 1934; Donovan, 1957). Large-scale trough and planar cross-stratification indicate north-east flowing currents.

Gule Horn Member. The Gule Horn Member is approximately 195 m thick along the west coast of Hurry Inlet where it overlies the Rævekløft Member (Surlyk *et al.*, 1973). To the north, in Lias Elv and Horsedal the member is approximately 200 m thick and directly overlies the Kap Stewart Formation. The sediments in Astarte Kløft and at Constable Pynt are characterised by tidal and shelf facies associations and have earlier been interpreted as offshore estuarine in origin (Sykes, 1974b). The tidal sediments are dominated by lenticular-, wavy- and flaser-bedded very fine to fine sandstone, alternating coarse sandstone and mudstone and cross-bedded medium- to coarse-grained sandstone. No indications of subaerial exposure were observed in the Gule Horn Member at Neill Klinter and probably the sediments were laid down on subtidal shoals. The shelf sediments are dominated by siltstone occasionally with storm sand layers. At the top of the succession at Constable Pynt possible tidal ridge deposits are present, suggesting that the palaeoshelf may have been both storm and tide dominated. Palaeocurrent data measured on small- and large-scale cross-stratification are strongly bipolar from N–S to NE–SW, but herringbone structures are seldom observed. The deposits of the Gule Horn Member at Lias Elv are very similar to the deposits along Neill Klinter. Palaeocurrent data from Lias Elv are also strongly bipolar from north-east and south-west. The crests of the wave ripples at Lias Elv are usually orientated ESE–WNW, possibly suggestive of an ESE–WNW trending coastline.

At Horsedal there seems to be a gradual transition between the Kap Stewart Formation and the Gule Horn Member. Wave-generated facies associations continue to dominate the deposits but bipolar palaeocurrent directions and current-generated wavy and flaser bedding indicate an increasing tidal influence in the Gule Horn Member. Shelf facies associations are not as common as to the south, and the local presence of rootlet horizons and thin coal seams indicate a more coastal position for the deposits in Horsedal. Palaeocurrent data are bipolar with currents from north-north-east and south-south-west. The crests of the wave ripples in Horsedal are orientated ESE–WSW pointing to the same palaeo-coastline orientation as in the Kap Stewart Formation.

Ostrea Elv Member. The tidal deposits of Gule Horn Member pass gradually throughout the area into the sediments of the Ostrea Elv Member. This member is approximately 90 m thick (Surlyk *et al.*, 1973) and is composed of very well sorted fine- to medium-grained sand. Bioturbation and the uniform grain size of the sediments often obscure primary sedimentary structures but locally wave ripples and large-scale cross-bedding are present. At the very top of the sequence at Horsedal a well exposed section with hummocky cross-stratification is present, which may indicate a storm-dominated shelf environment. The marine fauna present in several horizons of the Ostrea Elv Member is listed by Rosenkrantz (1934) and Donovan (1957).

The Neill Klinter Formation contains a wide variety of trace fossils which represent a wide range of behavioural patterns. A classification of behaviour gives at least three ethological groups (*domichnia, repichnia* and *fodichnia*), whereas the spatial distribution of the traces allows recognition of several trace fossil assemblages. These can possibly be followed throughout Jameson Land and southern Scoresby Land.

Laboratory work and future studies

The main objective for future work is the development of detailed and regional facies models for the sedimentary history of the Jameson Land basin during Late Triassic and Early Jurassic times. Detailed ichnofacies studies will be incorporated in the facies models. In order to evaluate the reservoir potential, porosity and permeability analyses will be carried out. Special emphasis will be placed on the correlation between reservoir properties and lithofacies. Palynological studies and source rock-analysis will be made on samples from shale sequences. Detailed biostratigraphical studies of the two formations in southern Scoresby Land are important as the precise age relations are very poorly known. Field work is planned to be continued in 1988 and 1989. In 1988 the main investigation will be in the areas along Hurry Inlet and Carlsberg Fjord near the eastern basin margin, in Schuchert Dal to the west, and around Gurreholm Bjerge to the north.

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A preliminary study of the porosity and permeability of limestones and sandstones in the Jameson Land area, central East Greenland

Lars Stemmerik

The Upper Palaeozoic – Mesozoic sequence in the Jameson Land area (figs 1, 2) is well known from numerous studies carried out since GGU initiated mapping of the area in 1968 (e.g. Clemmensen, 1980a, b; Surlyk *et al.*, 1973, 1984, 1986a; Surlyk, 1983, 1987; Heinberg & Birkelund, 1984). Based on these studies, Surlyk *et al.* (1986b) proposed that a number of Upper Permian – Cretaceous formations have a reservoir potential. Sandstone and limestone samples from these and a few additional formations (fig. 2) have been analysed for porosity and air-permeability. The analyses, all performed on surface samples or shallow cores, represent the first numerical approach to the evaluation of reservoir properties in the Jameson Land – Scoresby Land basin, and form a preliminary background for a more detailed reservoir study presently being undertaken by GGU.

The analyses were carried out commercially by the Core Analysis Laboratory of the Geological Survey of Denmark (Haslund, 1985; Springer, 1986). Air-permeability, He-porosity and grain density were measured conventionally on uncleaned 1" × 1" plugs after humidity drying.

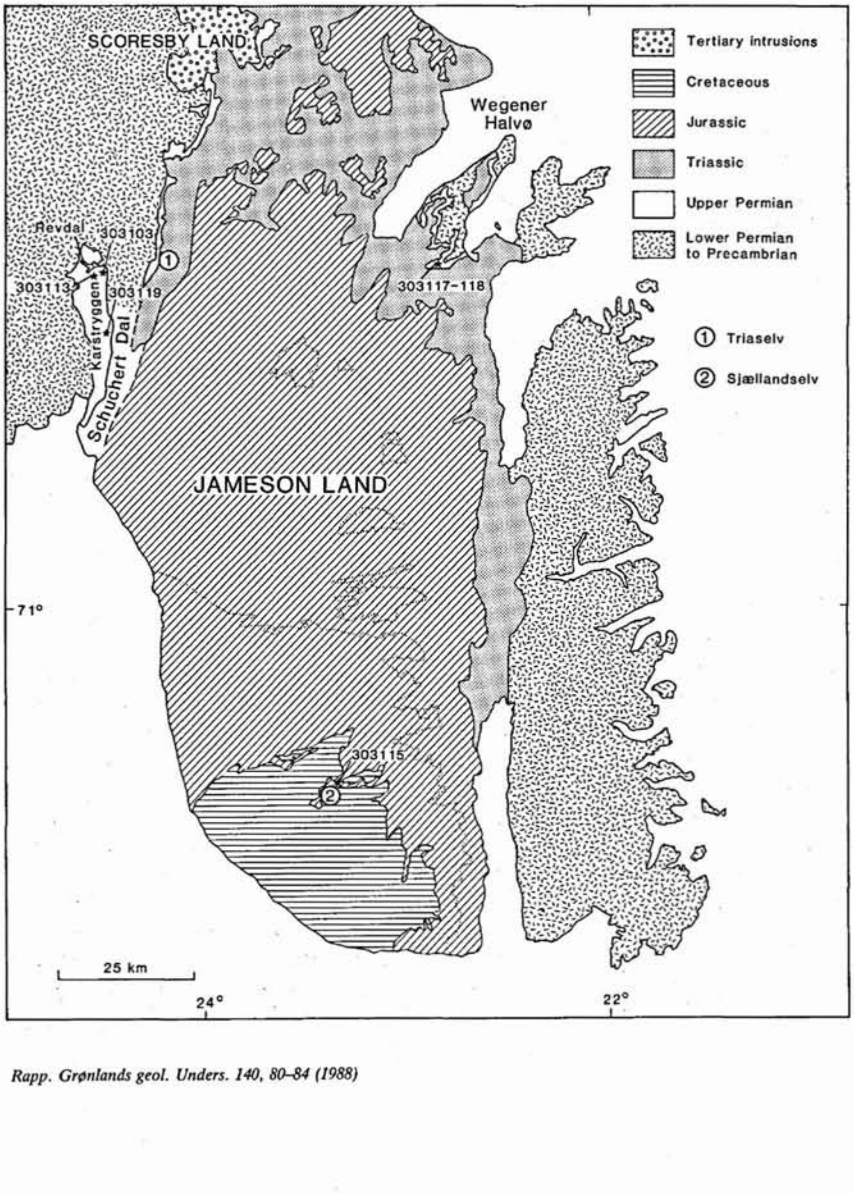


Fig. 1. Geological map of Jameson Land showing position of localities mentioned in the text and drilling locations. The geology is simplified from Surlyk *et al.* (1973).

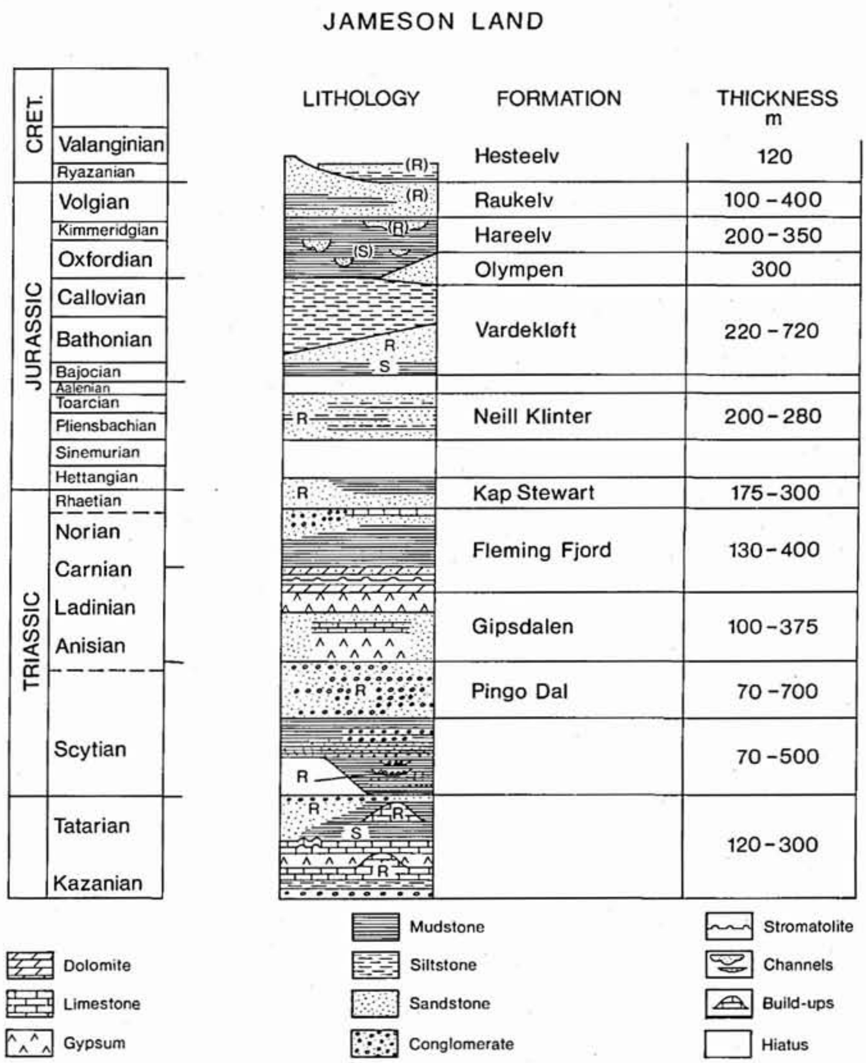
Upper Permian

Within the Upper Permian sequence potential reservoir rocks are confined to the limestones of the Karstryggen and Wegener Halvø Formations (Surlyk *et al.*, 1986a; Hurst *et al.*, in press) and the sandy part of the Schuchert Dal Formation (Surlyk *et al.*, 1986b). The juxtaposition of the Karstryggen and Wegener Halvø limestones with the potential source rocks of Ravnefjeld Formation makes these some of the more interesting targets in the subsurface. Therefore most of the analysed material comes from these two formations.

The Karstryggen Formation is 10–100 m thick and includes a variety of hypersaline limestones and calcium sulphate evaporites (Surlyk *et al.*, 1986a, b; Stemmerik, 1987). The analysed material includes 13 limestone samples from a shallow core (GGU 303101) drilled in the middle part of the formation near Revdal (fig. 1). The results of the analyses are shown in fig. 3a.

The Wegener Halvø Formation varies in thickness from less than 5 m centrally in the basin to more than 150 m along the margin (Surlyk *et al.*, 1986a). The thick platform sequences are dominated by algal mounds in the lower part, and bryozoan-cement mounds and shallow-water limestones in the upper part (Surlyk *et al.*, 1986a; Hurst *et al.*, in press). The analysed material includes 33 core samples from bryozoan-cement mounds at Wegener Halvø (GGU 303117, 303118) and Karstryggen (GGU 303113), and five samples from an algal mound at Karstryggen (core GGU 303119) (for location see fig. 1). The results of the analyses are shown in fig. 3b.

Fig. 2. Schematic stratigraphic section of the Upper Permian – Mesozoic sequence of Jameson Land. Main lithology and maximum thickness shown. R indicates potential reservoir and S indicates potential source rock. Modified from Surlyk *et al.* (1986b).



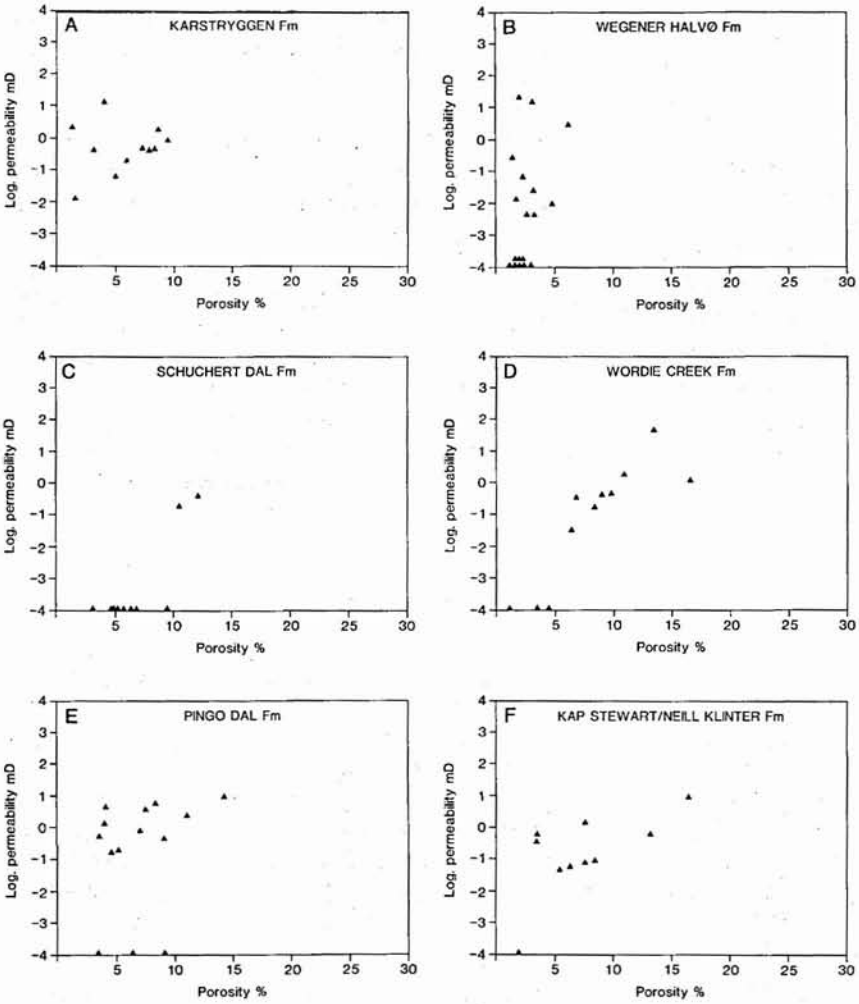


Fig. 3. Porosity–permeability plots of the Upper Permian – Lower Jurassic formations mentioned in the text.

Schuchert Dal Formation. The sandy part of the Schuchert Dal Formation, the Bredehorn Member, varies in thickness from 50–150 m. It consists of fine- to medium-grained micaceous sandstone deposited in short-headed fan deltas (Surlyk *et al.*, 1986a). The analyses were performed on surface samples from the east slope of Schuchert Dal (fig. 1). The results are shown in fig. 3c.

Triassic

Within the Triassic sequence it is suggested that potential reservoir rocks include sandstones from the Wordie Creek and Pingo Dal Formations.

Wordie Creek Formation. Potential reservoir rocks are confined to large submarine sandy channels within the otherwise shale-dominated unit. The largest channel sandstone recognized is more than 20 m thick and several kilometres across. Most of the analyses were carried out on surface samples from this sandstone unit in Triaselv (fig. 1). However, a few samples from thin sandstone beds elsewhere were also included. The results are shown in fig. 3d.

The Pingo Dal Formation varies in thickness from 70 m centrally in the basin to 700 m along the margins (Clemmensen, 1980a, b). The formation is dominated by two wedge-shaped members composed of alluvial-fan sandstones and conglomerates. The analysed material only includes surface samples from the northern part of the region. Sandstones from the two members have similar porosity/permeability values (fig. 3e).

Jurassic

The Jurassic succession outcrops in most of Jameson Land (fig. 1). Hence potential reservoirs are limited to the southern part where the Jurassic rocks are covered

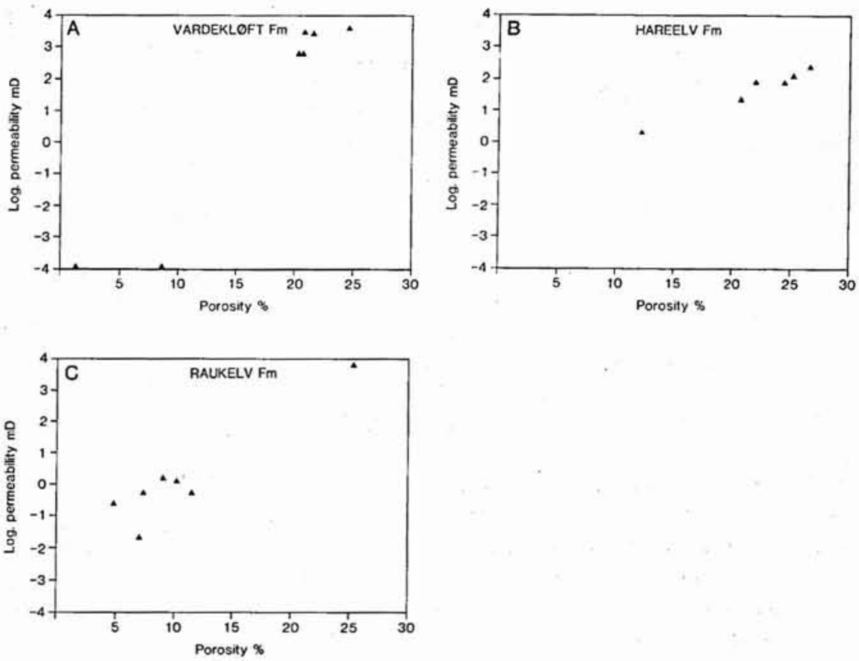


Fig. 4. Porosity–permeability plots of the Middle Jurassic – Lower Cretaceous formations mentioned in the text.

by Cretaceous sediments (fig. 1). However, the Lower Jurassic Kap Stewart and Neill Klinter Formations may be buried deeply enough to form potential reservoirs also in northern Jameson Land. In addition to the Lower Jurassic formations, the Middle Jurassic Vardekløft Formation and the Upper Jurassic Hareelv Formation include potential reservoir rocks. The late Jurassic succession is mainly of interest for possible future offshore exploration.

Kap Stewart and Neill Klinter Formations. The material available from these formations is sparse and relatively poor and only includes surface samples from the south eastern margin of the basin. Here, the Kap Stewart Formation is 200 m thick and composed of fluviatile sandstones. The Neill Klinter Formation is 250–300 m thick and dominated by tidal deposits. Analyses from the two formations are presented together in fig. 3f.

Vardekløft Formation. Within this formation potential reservoir rocks are found in the sandy Pelion Member. This member consists of a large wedge-shaped body of medium- to coarse-grained shelf sandstone, which increases in thickness northwards from 10 m to more than 600 m (Surlyk *et al.*, 1973; Heinberg & Birkelund, 1984). The analyses were carried out on surface samples from the northern part of the region. The results are shown in fig. 4a.

Hareelv Formation. Potential reservoir rocks in the Hareelv Formation are restricted to slope and deep-shelf gully sandstones enclosed in the otherwise shale-dominated unit (Surlyk, 1987). The sand bodies are up to 50 m thick and hundreds of metres wide, and may be more than 5 km long in downcurrent direction (Surlyk, 1987). Samples from core GGU 303115 drilled in Sjællandselv (fig. 1) were analysed. The results are shown in fig. 4b.

Cretaceous

Both the Raukelv and the Hesteelv Formations include sandstones that may be potential reservoirs. However, the distribution of the Cretaceous sediments in Jameson Land (fig. 1) prevents them being reservoirs here, so they are mainly of interest for possible future offshore exploration. During this study only material from the Raukelv Formation was available.

The Raukelv Formation is estimated to be 300 m thick (Surlyk *et al.*, 1973). It consists of thick, homogeneous or large-scale cross-bedded sandstone units alternating with shaly siltstones. The materials analysed are surface samples of sandstones from Sjællandselv (fig. 1). The results are shown in fig. 4c.

Discussion

The use of surface material to predict reservoir properties in the subsurface is very difficult, particularly where carbonates are involved. The main problem is the timing of different diagenetic events in relation to hydrocarbon generation and migration.

It is evident from figs 3a and 3b that the analysed Upper Permian limestones have very low values for both porosity and permeability. However, diagenetic studies show that much of the primary porosity is filled by a late high-temperature cement (Surlyk *et al.*, 1986a; Hurst *et al.,* in press). The reservoir potential in the subsurface is therefore strongly dependent on the age of this cement relative to hydrocarbon migration.

The sandstones show highly variable porosities and permeabilities, both from formation to formation, and within a single formation. It is evident that the most promising analytical values are recorded from the Vardekløft and Hareelv Formations (figs 4a & 4b). Also it is obvious that the porosities and permeabilities are generally higher in the Jurassic and Cretaceous sandstones compared with the Upper Permian and Triassic values (figs 3, 4). However, detailed diagenetic studies are needed to see whether this pattern is related to differences in the primary lithology, thermal maturity and subsidence depth, or if it is related to preferential secondary leaching of carbonate cement in the Jurassic – Cretaceous sandstones during uplift. Also, the Triassic and Upper Permian feldspar-rich sandstones may have increased secondary porosity in their deeper buried settings in the subsurface.

This preliminary study of potential reservoir rocks will be followed by more detailed diagenetic studies of the Upper Permian Karstryggen and Wegener Halvø Formations, the Lower Jurassic Kap Stewart and Neill Klinter Formations (see Dam, 1988), and the Lower Cretaceous Raukelv Formation.

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Preliminary report of fission track studies in the Jameson Land basin, East Greenland

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Fission track (FT) analysis is especially suited to reveal and date low temperature events. The closure temperature of apatite (100 ± 30°C) and its annealing characteristics in the interval of 70–125°C are especially relevant to the study of the maturation of hydrocarbons (Gleadow *et al.*, 1983).

FT analyses were made on Permian to Cretaceous, quartzose sandstones and arkoses from the Jameson Land basin. Both FT ages and track length distributions for apatites were obtained for samples taken along the western and eastern margin of the basin (fig. 1 and Table 1) in order to study the tectonic and thermal history of the area. The investigation takes advantage of earlier FT work in the neighbouring Caledonian mountain belt which is believed to be the source of the terrigenous material, including the apatites, which make up the sediments (Hansen, 1985). A report of further investigations in this area is in preparation.

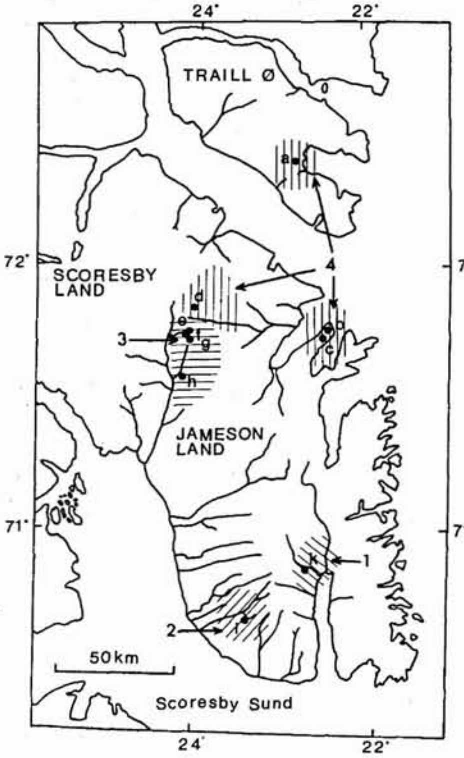

Fig. 1. Sample localities, Jameson Land, East Greenland. Shadings indicate different thermal type areas. Key to localities in Table 1.

Technique

Samples of *c.* half a kilogram were crushed and separated using magnetic and heavy liquid methods. Most samples yielded apatites. Polished and etched apatite mounts were irradiated together with mica detectors in the J1 facility of the HERALD reactor in Aldermaston, England, for age determinations. Specially polished and etched mounts were prepared for track length measurements. The measurements were carried out using a Zeiss-Jena microscope with a × 100 oil objective and a cover slip, and either a calibrated net or a scale bar inserted in the × 12.5 ocular. The calibration used in the age determinations follows the suggestions of Hurford & Green (1983). The NBS SRM612 and Corning CN1 and CN2 glasses were used as fluence monitors and the Fish Canyon and Mt. Dromedary apatites as age standards in the zeta calibration.

Results

Table 2 shows results of FT apatite age determinations and mean track length distributions. Table 3 shows track densities for glass standards for the two irradia-

Table 1. Basic data for sample localities, Jameson Land, East Greenland

Locality	area type	GGU sample no.	elevation m a.s.l.	sediment formation age
Traill Ø				
a	4	327371	880	Cretaceous
Wegener Halvø				
b	4	298343	490	Triassic
c	4	327546	880	Permian
West Jameson Land				
d	4	221247	1100	Upper Permian
e	3	292005	900	Upper Permian
f	3	248628	400	Lower Permian
g	3	293284	771	Upper Permian
h	3	293252	213	Lower Triassic
South Jameson Land				
i	2	292016	300	Upper Jurassic
Ugle Elv, east Jameson Land				
k	1	248652	500	Upper Jurassic

Deposition ages are obtained from Piasecki (personal communication, 1987).

Table 2. Mean track lengths and apatite fission track ages of sedimentary rocks from the Jameson Land basin, East Greenland

Sample GGU no.	Irr. no.	$\varrho_s \times 10^5$ (N_s)	$\varrho_i \times 10^5$ (N_i)	grain no.	age(Ma) ± 1σ	mean length (μm) ± 1σ(N)
Traill Ø						
327371	KH18	7.4369 (160)	51.500 (1108)	7	15.57 ±1.69	†
Wegener Halvø						
298343				†	†	12.28 ±3.34 (44)
327546	KH18	5.2455 (86)	29.643 (486)	9	19.23 ±2.59	10.64 ±3.76 (33)
West Jameson Land						
221247	KH18	7.4746 (31)	40.025 (166)	4	20.29 ±4.20	10.83 ±3.20 (28)
292005	KH17	57.029 (340)	50.320 (300)	8	135.56 ±14.01	10.01 ±2.49 (37)
248628*	KH17	25.658 (281)	45.290 (496)	8	70.23 ±9.49	10.55 ±2.95 (110)
293284	KH18	8.2379 (71)	8.0058 (69)	7	111.10 ±20.19	†
293252	KH17	29.628 (432)	45.677 (666)	9	77.93 ±7.05	10.05 ±2.75 (100)
South Jameson Land						
292016	KH17	15.630 (472)	34.406 (1039)	15	54.68 ±4.71	11.51 ±2.30 (100)
Ugle Elv, east Jameson Land						
248652*	KH18	29.176 (467)	16.931 (271)	10	205.30 ±31.39	†

* failed the chi-squared test. Ages calculated on the basis of individual grain ϱ_s/ϱ_i ratios.
† not determined.
ϱ_s and ϱ_i are track densities of spontaneous and induced tracks respectively, N_s and N_i similarly are the number of tracks counted and N represents the number of tracks measured.

tions and the zeta values which were employed. Fig. 2 shows the measured track length distributions.

Discussion

Apatite FT ages, except for the area in the south-east which is described below, are annealing ages modified by a post-sedimentary heating event (Table 2). This is shown by FT ages which are younger than the 150 Ma apatite cooling ages found for the neighbouring basement (Hansen, 1985) and also younger than the sediment deposition age of individual samples given in Table 1. The FT age distribution suggests that different parts of the area suffered different heating histories. The different parts of the area are shown in fig. 1 and are described below.

(1) Area type 1 is an area in the south-east which is only weakly annealed and probably slowly uplifted. It gives an apatite FT age of *c.* 200 Ma. The material for the sample representing this area may originate from the nearby Hurry Inlet granite of 425 Ma (Hansen & Steiger, 1971). The FT age for the granite yields 203 Ma for apatite for a surface sample (Gleadow & Brooks, 1979) taken to represent apatite FT ages at today's erosion level for the Hurry Inlet granite. This means that apatites in the sediment may range from 425 to 200 Ma for an unannealed sample. The actual age spread found for the sediment (Upper Jurassic) is between 80 and 330 Ma, i.e. lower than expected for unannealed samples, the youngest being younger than the sediment age indicating that the partial annealing was experienced after deposition. The mild annealing suggests that the temperature was close to the lower temperature limit of the annealing interval, e.g. *c.* 80°C, which is in accordance with vitrinite reflectance measurements found for the area (E. Thomsen, personal communication, 1987).

(2) Area type 2 is an area in the south which was probably almost totally annealed (close to 125°C) and was slowly uplifted at, or later than 55 Ma ago, the thermal activity perhaps being indirectly related to the Scoresby Sund igneous activity at 53–56 Ma (Watt *et al.*, 1986). A

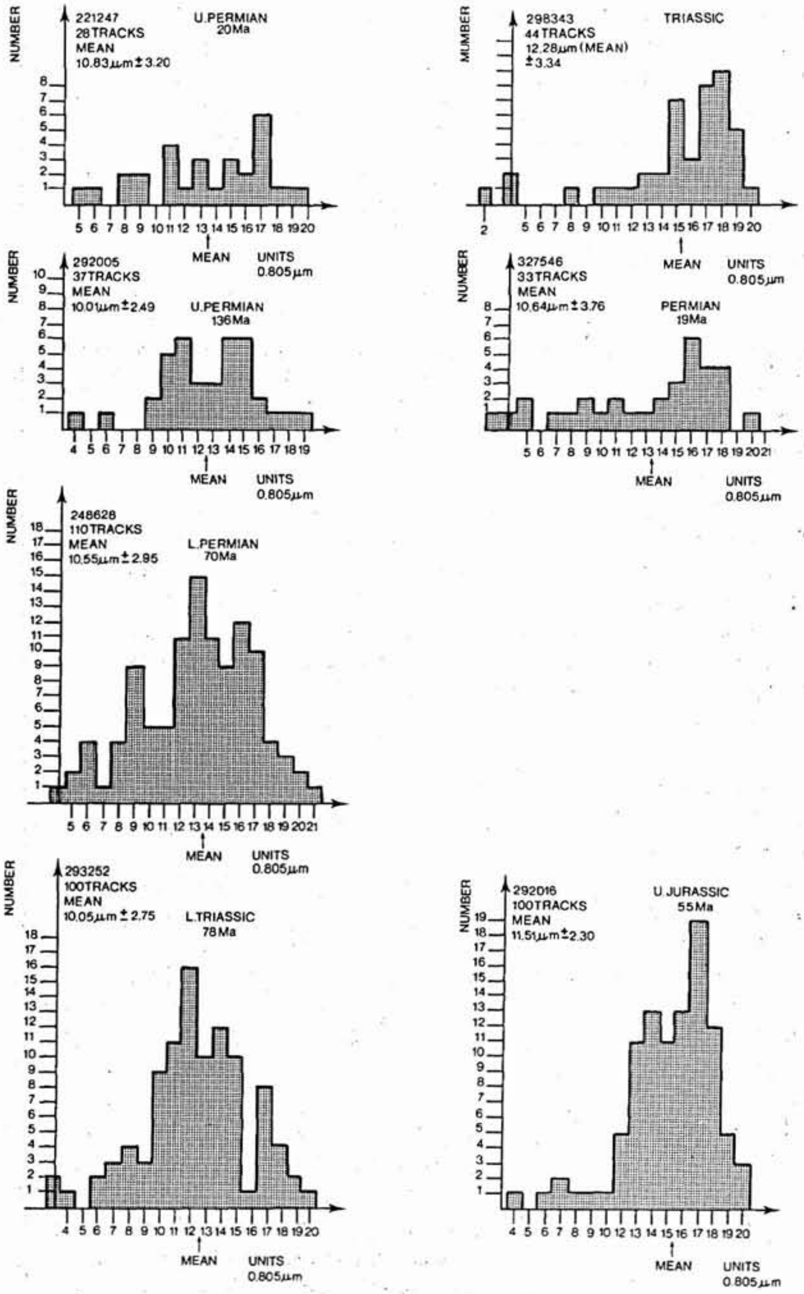

Fig. 2. Track length distributions, Jameson Land, East Greenland. Mean track lengths, uncertainties, deposition ages and apatite fission track ages are also given in the figure.

temperature not much lower than 125°C fits the maturity pattern revealed by, e.g. reflectance measurements (E. Thomsen, personal communication, 1987). This area could be related to the partly annealed area of type 3, representing a more deeply buried section.

(3) Area type 3 in the west shows varying ages, 70–135 Ma, which can be partly ascribed to different degrees of annealing due to different elevations (Table 1). Thus the samples from the highest levels cooled below the annealing interval first and also did not proceed so far into the annealing interval. Temperature differences at 30°C/km would be *c.* 15–20°C for differences in altitudes of *c.* 600 m which could produce marked age differences at temperatures around 100°C (Gleadow *et al.*, 1983). The time of the uplift may coincide with the uplift of the southern area. Hydrocarbon maturity measurements reveal a complex pattern with a major maturity shift between Lower and Upper Permian (Surlyk *et al.*, 1986) not found in the FT data.

(4) *Area type 4* in the north shows young uplift ages of *c.* 20 Ma which implie almost total annealing. The connection between the eastern and western part is unclear, as is the influence of the Werner Bjerge intrusion which also gives an apatite uplift age of 20.5 Ma (Gleadow & Brooks, 1979) and similar K-Ar ages (Schassberger, personal communication to C. K. Brooks). Maturity measurements also show these northern areas to be post-mature with respect to oil generation (Surlyk *et al.*, 1986).

The above findings are supported by track length distributions, mean track lengths, and their uncertainties (Table 2 and fig. 2). Heating into the annealing interval reduces track lengths, the track length distribution thus being diagnostic of the path through the heating interval (Gleadow *et al.*, 1986).

Track length distributions have so far been obtained for samples from the southern area (type 2), the western area (type 3), and the northern areas (type 4), although the restricted amount of material makes it difficult to obtain reliable results. However, the internal consistency of the few data allows a generalised description of track length distributions for three of the areas described.

The youngest area (type 4) shows the simplest track length distributions in its eastern part, where distributions are dominated by young tracks in a typical skewed uplift pattern (Gleadow *et al.*, 1986). However, the broad distribution and the low mean indicate a small admixture of pre-uplift age to the measured age. This means, as suggested above, that temperatures close to total annealing (125°C) were attained together with late uplift (after 20 Ma ago). The distribution for the sample taken close to the contact of Werner Bjerge intrusion (221247) does not reveal a typical uplift pattern. This may, however, be due to the difficult and restricted material and interpretation must await further determinations.

The track length distribution of the northern samples (area type 4) contrasts with the distribution for the western area (type 3) which shows a very broad symmetrical pattern. A symmetrical distribution is characteristic of partial annealing with only a small proportion of newly formed long tracks, i.e. the resulting mixed age is dominated by partly annealed old tracks and the uplift age is much younger than the mixed age. A temperature close to 100°C must be assumed as the maximum temperature experienced.

In the southern area (type 2) the track length distribution is narrower with a younger (55 Ma) age compared to the western area, but it still has an old component as shown by mean lengths and uncertainties; it is probably a result of further annealing due to deeper burial than found in the western area. This implies a temperature close to, but not exceeding, 125°C and a slow uplift, not much later than 55 Ma ago. The area has been uplifted together with the neighbouring areas which were never so deeply buried, as shown by their higher ages and greater spread in track lengths.

Table 3. Track density of glass standards, calibration factors

	ϱ_dSRM612 Number	ϱ_dCN1 Number	ϱ_dCN2 Number
KH17	8.089194 × 10^5 5242	24.86787 × 10^5 6446	22.96979 × 10^5 5954
KH18	7.100035 × 10^5 4601	21.90888 × 10^5 5679	21.64268 × 10^5 5610
ζ	288 ± (9.38%)	101 ± (4.95%)	105 ± (4.76%)

KH17 and KH18 are irradiation numbers, ϱ_d track densities of mica detectors for the respective glass standards and zeta calibration factors (Hurford & Green 1983) used for the age determination based on age standards.

Conclusions

In general the results support the maturity pattern found for the area by Surlyk *et al.* (1986). Assuming a geothermal gradient of 30°C/km, the maximum burial is constrained to be less than 3 km by by the maximum temperature experienced (less than *c.* 125°C), except perhaps to the north. During uplift, sediments in the southern part first passed the 100°C isotherm not much later than 55 Ma ago and were uplifted together with the less deeply buried neighbouring areas. To the north, uplift led to the passing of the 100°C isotherm much later, around 20 Ma ago. The reasons for the young ages to the north, compared to more southern areas could be delayed uplift followed by a rapid uplift during the last 20 Ma. The rapid uplift is not likely to have proceeded for a longer time span than the 20 Ma as this would lead to exposure of sedimentary rocks of higher formation age to the north compared to the rest of the area, and this did not occur. Alternatively, younger ages could also result from a higher geothermal gradient which would be likely to accompany the Tertiary intrusive activity. Finally a combination of the two explanations is possible.

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Studies of the onshore hydrocarbon potential in East Greenland 1986–87: field work from 73° to 76°N

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The study of the Devonian to Cretaceous sequence in central and northern East Greenland was continued in 1987. Field work was carried out from early July to mid August covering the region between Ymer Ø and Hochstetter Forland (fig. 1). This was the second year of a two-year field work programme (Marcussen *et al.*, 1987) which forms part of a regional programme comprising sedimentological, stratigraphic, structural, and petroleum geological studies of the sedimentary basin in central East Greenland (e.g. Surlyk, 1983; Surlyk *et al.*, 1984, 1986a,b).

Stordal in Hudson Land, which offers a natural landing strip for STOL aircraft, was used as base camp for the 1987 expedition. The expedition group of 15 included four to five geological field parties and five supporting personnel, including a helicopter pilot and mechanic. In addition a five-man British-Danish East Greenland 'vertebrate–paleontological' expedition (Bendix-Almgreen, 1988) and a group from GGU planning the 1988 expedition to North-East Greenland were present. Two teams (led by P.-H. Larsen and H. Olsen) studied the tectonics and sedimentology of the Devonian succession. Two teams (led by L. Stemmerik and S. Piasecki) investigated the Carboniferous, Permian and Triassic sedimentology of the region. The Jurassic and Cretaceous sequences were studied by two teams (led by S. Piasecki and in the late half of the season by H. Nøhr-Hansen). All the teams collected material for source rock analyses and a large number of samples were also collected for determining reservoir rock properties.

Devonian

Sedimentological studies. The field work in 1987 was concentrated on Ymer Ø, Gauss Halvø, Moskusokselandet and Strindberg Land (figs 1, 2). Detailed facies analyses were made by vertical facies logging and 2-

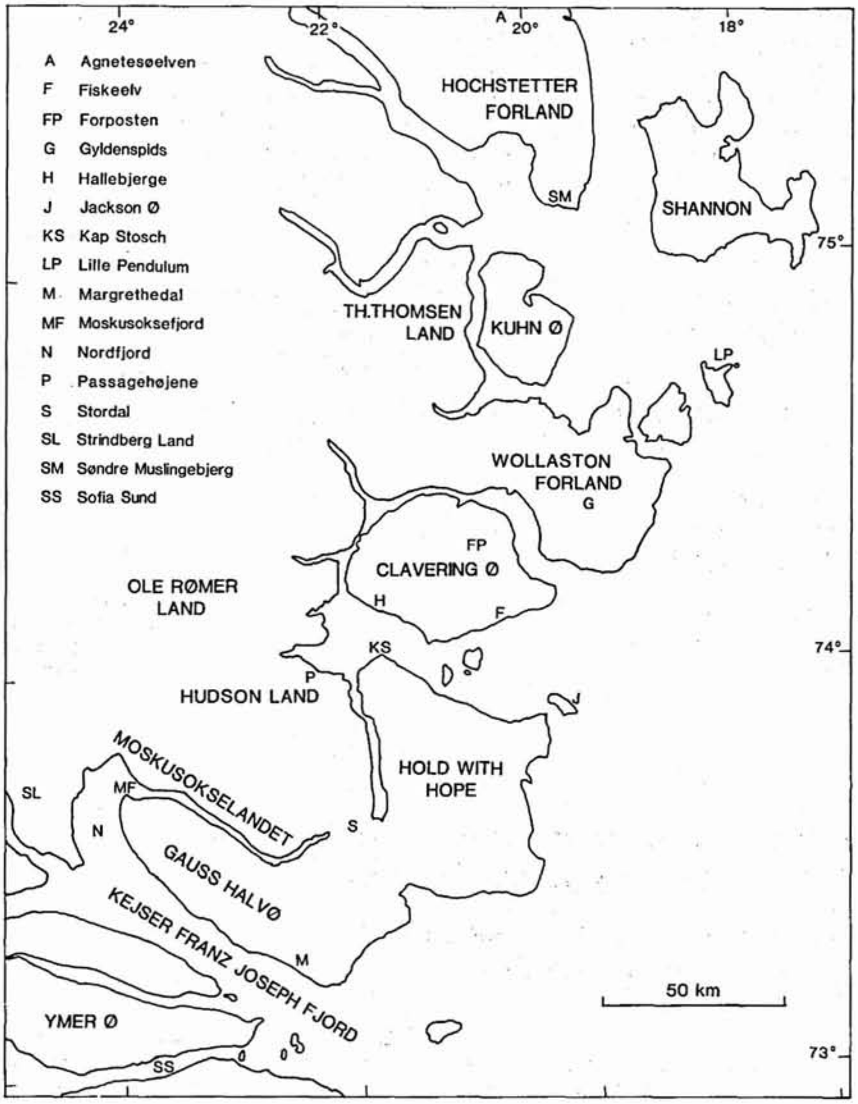

Fig. 1. The study area with localities mentioned in text.

dimensional facies mapping in the Middle to Upper Devonian Kap Kolthoff Supergroup (Friend *et al.*, 1983). Sandy braidplain deposits, alluvial fan conglomerates and windblown sandstones were present. The Upper Devonian Kap Graah Group (Friend *et al.*, 1983) was studied in detail in Ymer Ø and Gauss Halvø. By facies association mapping, a stratigraphic subdivision of the formerly undivided group was attempted. Facies analysis of meandering stream deposits implies an asymmetric development of the point bar deposits which may be controlled by tectonics. Similarly an asymmetric distribution of facies associations and palaeocurrent directions seemed to indicate a tectonic influence on the depositional environments. The Upper Devonian Mt. Celcius Supergroup (Friend *et al.*, 1983) was briefly studied on Stensiö Bjerg, Gauss Halvø. Facies associations were logged and two black shale units of potential source quality, 1 m and 5 m thick, were sampled in detail for analytical work.

Structural studies. The structural investigations in the Devonian outcrop area were concentrated on Moskusokselandet, Ole Rømer Land, Hudson Land and Ymer Ø (figs 1, 2). Selected areas and meso-structures were mapped on aerial photographs (1:50 000), while larger structures were studied by reconnaissance work supported by helicopter.

The coastal cliffs of Sofia Sund, Kejser Franz Joseph Fjord, Nordfjord and Moskusoksefjord were photographed by Jacob Lautrup. These coast panoramas together with new aerial photographs taken by the Geodetic Institute in 1987 (1:150 000) will be important

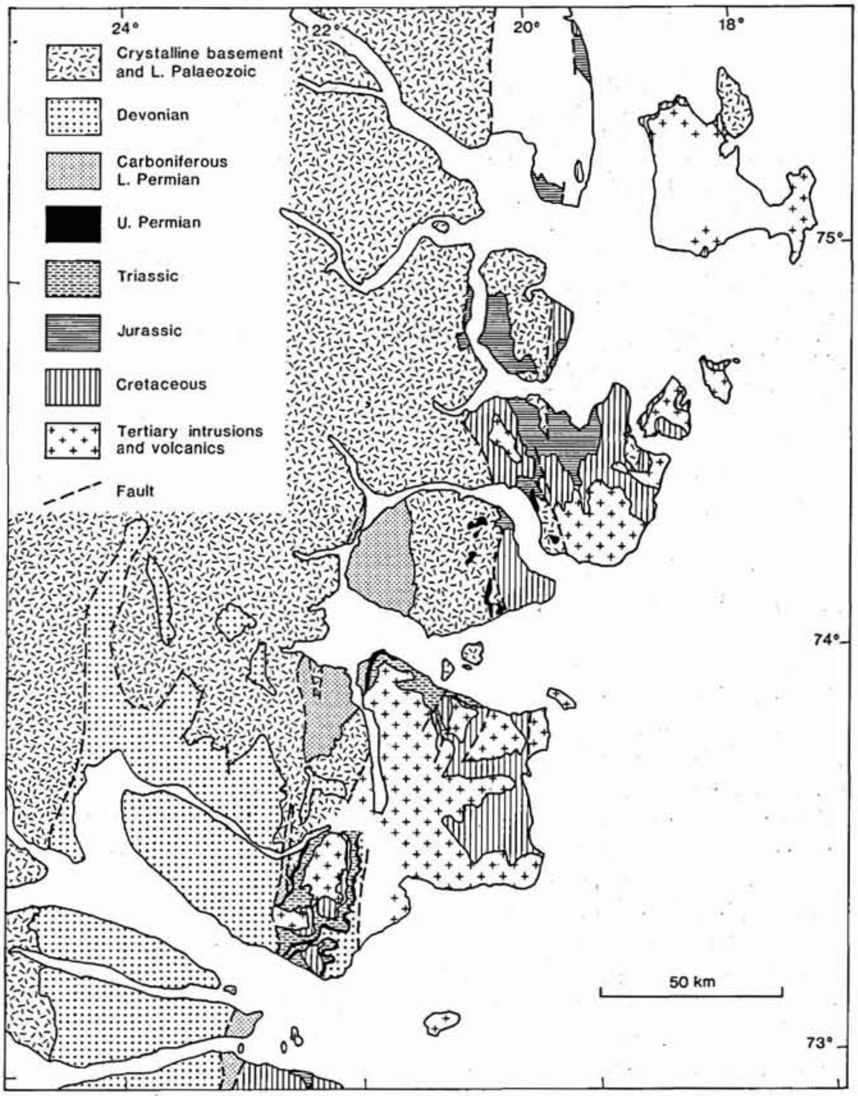

Fig. 2. Geological map of the study area. Simplified after Koch & Haller (1971) with minor revisions based on Surlyk (1978b) and 1987 field work.

tools in the continued structural work. According to Bütler (1959) the structural style is dominated by thrusting and folding in various synsedimentary phases, the Hudson Land phases, resulting in angular disconformities and breaks in the sedimentation. The present work has shown that these deformations are local in character reflecting the sedimentological evolution in local parts of the basin, but not necessarily in the entire basin. In Carboniferous time the area was deformed during the Ymer Ø phase of Bütler (1959). This deformation especially affected the southern part of the basin and the 1987 work shows that the deformation here caused the development of low-angle thrusts and ramp anticlines in a contractional thin-skinned tectonic environment (fig. 3). Some attention was paid to the granitic 'Moskusokselandet Inlier' which Bütler (1959) suggests was developed during the Devonian. Eleven samples were collected for radiometric dating (Rb/Sr) to investigate the timing of the granitic emplacement within the inlier. Collection of fossil plant material from the Devonian sandstones was continued this year. The state of preservation is rather poor and preliminary investigation of the material by Dianne Edwards, University College, Cardiff, Wales, has not yet yielded any new stratigraphic information.

Carboniferous

Sediments of Carboniferous age were studied briefly around Moskusoksefjord and Passagehøjene, and in some detail in Hallebjerge, western Clavering Ø (figs 1, 2). Tentatively, this more than 600 m thick succession is

Fig. 3. Devonian sandstone forming a huge box-fold which to the east is overridden by a thrust. Rødebjerg at the south coast of Ymer Ø. Height of cliff is approximately 1600 m.

correlated with the lowermost of the three units described from the Traill Ø – Ymer Ø area by Marcussen *et al.* (1987).

The succession is dominated by fluviatile sandstones. Malmquist (1932) and Säve-Söderbergh (1934) subdivided the succession at Hallebjerge and Passagehøjene, respectively, based on the different colour of the sandstones and the content of the conglomerates.

The present study indicates that fine-grained, shaly deposits are more widespread than previously assumed. The recovery at western Clavering Ø of two shaly sequences in the lower red sandstone unit of Malmquist (1932) may be of interest as potential source rocks. The 15–20 m thick shaly sequences yield abundant ostracods and are thought to be of lacustrine origin. Lacustrine shales of Carboniferous age with a fair to good source-rock potential are known from several locations further to the south and suggest that the Clavering Ø shales may hold similar source-rock quantities.

Upper Permian – Lower Triassic

The outcrops of Upper Permian – Lower Triassic sediments in the northern part of the basin are geographically subdivided into three regions: Gauss Halvø, Kap Stosch and eastern Clavering Ø. (figs 1, 2). These regions were investigated in detail for sedimentology, stratigraphy, thermal maturity and source-rock analyses. The lithological subdivision of the Upper Permian sequence, the Foldvik Creek Group, employed by Surlyk *et al.* (1986a) in the Jameson Land region, appears to be applicable to the complete region with only minor adjustments.

The basal Upper Permian Huledal Formation is deposited unconformably on a variety of substrata including crystalline basement, post-Caledonian intrusions, Devonian and Carboniferous sediments. The thickness depends on local topography and differentiated subsidence of the substratum and varies from 0 to more than 100 m. The overlying Karstryggen Formation is 5 to more than 30 m thick and consists of a variety of hypersaline shallow marine limestones and evaporites. Preliminary investigations suggest that deposition took place in environments similar to those described from the formation in Jameson Land (Surlyk *et al.*, 1986a).

The Karstryggen Formation is overlain by normal marine limestone of the Wegener Halvø Formation along the western margin of the basin, particularly in north-eastern Clavering Ø, and by scattered outcrops in southern Wollaston Forland. The formation is mainly composed of bedded biogene limestone of probable reef flank origin. However, *in situ* reefs or mounds were not found in this less than 60 m thick carbonate sequence. Laterally to the east, shales of the Ravnefjeld Formation were deposited.

The Ravnefjeld Formation, formerly Posidonia Shale, occurs throughout the region. The maximum thickness appears to be 15 to 20 m. Maync (1942) indicated 70 m of 'Posidonia Shale' from Fiskeelv on Clavering Ø, but due to scree cover of the section, no precise measurement of the thickness is possible; however, the estimated thickness is about 20 m rather than the 70 m suggested by Maync (1942). The Ravnefjeld Formation is generally reduced in thickness to a few metres near carbonate buildups at the western margin of the basin. The Ravnefjeld Formation may be subdivided into a lower grey, bioturbated siltstone with calcareous beds and an upper black, laminated silty shale with intervals of laminated, graded or massive carbonate beds and high concentrations of *Posidonia permica*. The lower unit is generally less than 5 m in thickness but exceptionally a thickness of 25 m was measured at the north side of Forposten, Clavering Ø. This lower unit is not known in the outcrops in Jameson Land. However, the upper part, the traditional 'Posidonia Shale', is very similar in the northern region to the outcrops in Jameson Land except for the presence of a rich nektonic vertebrate fauna which is especially confined to the calcareous intervals (Nielsen, 1935).

The overlying shale of the Schuchert Dal Formation is relatively calcareous compared to the southern region. The shale, tentatively included in the Oksedal Member, may be subdivided into a lower unit of calcareous silty shale and/or massive carbonate beds (formerly 'Martinia Limestone') and an upper unit of grey, green to red calcareous siltstone which consistently forms the top of the Upper Permian sequence. Carbonate debris flows (formerly 'Productus Limestone') rich in brachiopods,

Fig. 4. Erosive channels filled with giant cross-bedded sandstone and conglomerates. Wordie Creek Formation, Kap Stosch. Height of lower sandstone unit is approximately 25 m.

bryozoans and gastropods, occur scattered throughout.

The Lower Triassic Wordie Creek Formation follows the Upper Permian succession with a sharp boundary most likely representing a depositional break. However, in most cases, Triassic sediments conformably overlie the Permian sequence without any significant erosion in between. The earliest Triassic sediments are siltstone and conglomerates with coarse- to fine-grained sandstone. Higher in the sequence fine-grained wavy-bedded to lenticular sandstones alternate with red, brown and green siltstone. Six horizons of sandstone in the upper silty sequence transform laterally into deep erosional channels filled with sandstone and conglomerates. Giant epsilon cross bedding (fig. 4) can be seen within these channels.

Jurassic–Cretaceous

The Jurassic–Cretaceous succession is widely exposed particularly in the eastern part of the area (fig. 2). It overlies unconformably Caledonian to Triassic rocks and is covered by Tertiary plateau basalts (fig. 2).

The aim of the field work in the Jurassic–Cretaceous sequence was to obtain samples of material for stratigraphical, source potential and maturity studies. In order to understand the regional maturity pattern, sampling in large areas in the eastern part of the region, including Hold with Hope, Clavering Ø, Lille Pendulum, Shannon, Hochstetter Forland (figs 1, 2) was carried out during a helicopter reconnaissance programme.

The Middle Jurassic Vardekløft Formation was studied in some detail in Hochstetter Forland, Wollaston Forland and Th. Thomsen Land (fig. 1). Material from the type section of the Muslingebjerg Member (Surlyk, 1977) at Søndre Muslingebjerg (fig. 1) and a previously poorly known section of the Pelion Member at Agnetesøelven (fig. 1) may be particularly important in understanding the Middle Jurassic stratigraphy.

The overlying dark silty shale of the Upper Jurassic Bernbjerg Formation and the syntectonic sediments of the Upper Jurassic – Lower Cretaceous Wollaston Forland Group (Surlyk, 1978a) were sampled intensively.

The Mesozoic succession is terminated by a more than 500 m thick transgressive sequence of dark silty shale with streaked and cross-bedded sandy horizons in the upper part. The shales overlie unconformably Upper Permian – Lower Cretaceous sediments. They are poorly dated as Aptian – Albian on the basis of very scattered macrofossils (Maync, 1949). Therefore, a systematic palynological sampling programme was carried out to establish a biostratigraphical zonation. However, Tertiary sills preferentially intrude these shales and may have altered the organic material in much of the area. It is hoped that analyses will show that well preserved sections are found on the north and east coast of Hold with Hope, Jackson Ø, Clavering Ø, Kuhn Ø and several places in Wollaston Forland.

The thickness of the Albian – Aptian sequence varies from 25 to 115 m in Kuhn Ø to more than 500 m at the mountain Gyldenspids (fig. 1) in Wollaston Forland. Further to the south on Hold with Hope and Gauss Halvø the thickness is approximately 400 m.

Fig. 5. Bitumen filled vugs in an Upper Permian algal boundstone, Margrethedal.

Evidence of hydrocarbon generation

Previously, evidence of hydrocarbon generation was presented by Secher & Steenfelt (1976), who found degraded bitumen in Devonian and Carboniferous rocks from geographically widely separated localities in the south-western part of the study area. Many of the localities mentioned were revisited this summer. The bitumen occurrences are very localized and seem to be related to hydrocarbon generation in the vicinity of major faults or intrusions.

New evidence of hydrocarbon generation includes:

(1) A local discovery of degraded bitumen in an Upper Permian oolitic grainstone at Kap Stosch.

(2) Common occurrences of bitumen in Upper Permian algal boundstone (fig. 5) and Triassic sandstone at Margrethedal. At this locality, the bitumen is associated with hydrothermal minerals such as flourite, dolomite and galena in the limestone. Generation of hydrocarbons probably took place in the Ravnefjeld Formation shales during intrusions of the numerous Tertiary sills in the area.

Discussion

The investigations in the Devonian mainly focussed on the tectonic and sedimentary history of the basin. It was evident from the 1986 field work (Marcussen *et al.*, 1987) and the following analyses that the source potential of the Devonian sequence is very restricted. This was confirmed as only two very thin shaly sequences of restricted lateral extension were found.

The oil potential of the onshore part of the East Greenland basin is apparently confined to the Carboniferous to Cretaceous part of the sequence. Rock Eval analyses of parts of the material collected this year indicate that the potential source rock units identified in the southern part of the area (Surlyk *et al.*, 1986a; Piasecki, 1986, 1987; Marcussen *et al.*, 1987) also have source potential. This is particularly the case for the marine shales of the Upper Permian Ravnefjeld Formation and the Upper Jurassic Bernbjerg Formation. In contrast the Carboniferous lacustrine shales and the thick Cretaceous Aptian – Albian shale sequence have not yet proved to have any source potential.

Potential reservoir units occur throughout the sequence. However, onshore prospects may be found primarily in the Carboniferous to Triassic succession having fluviatile sandstones, carbonate mounds or submarine channel sandstones in stratigraphical positions close to the respective potential source rocks.

Acknowledgements. The Royal Danish Air Force provided transportation to and from Mestersvig, coordinated by Major J. F. Clemmensen. The landing strip at Mesters Vig is kept open by Nordisk Mineselskab on behalf of the Mineral Resources Administration for Greenland. Nordisk Mineselskab also provided facilities and help during mobilization and demobilization in Mestersvig. Flugfélag Norðurlands hf., Island, transported personnel and goods from Mestersvig to Stordal. Unifly, Denmark, provided helicopter transportation during the season with L. Rahm and B. Hvidberg as pilots and K.

Kristiansen as mechanic. I. K. Olsen and J. Lautrup worked as base personnel while J. Bojesen-Koefoed, J. Halskov, G. S. Nielsen, P. Schiøler and K. Zinck Jørgensen assisted in the field and at base camp. H. Olsen's sedimentological studies in the Devonian are supported by the Carlsberg Foundation. M. Larsen and B. Sikker Hansen helped to prepare the manuscript.

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Upper Devonian and Upper Permian vertebrates collected in 1987 around Kejser Franz Joseph Fjord, central East Greenland

Svend Erik Bendix-Almgreen, Jennifer Alice Clack and Henrik Olsen

In July and August 1987 a five-man British/Danish vertebrate palaeontological expedition carried out field work on Gauss Halvø and eastern Ymer Ø to collect fossil vertebrates from exposures of the Upper Devonian continental and Upper Permian marine deposits. In the context of sampling for vertebrates the Upper Devonian deposits were last visited in 1955 when one of the writers (SEB-A) assisted the late Dr Eigil Nielsen with field work while the Upper Permian marine deposits have not been examined since Bendix-Almgreen's field work in the Kap Stosch area in the summer of 1967.

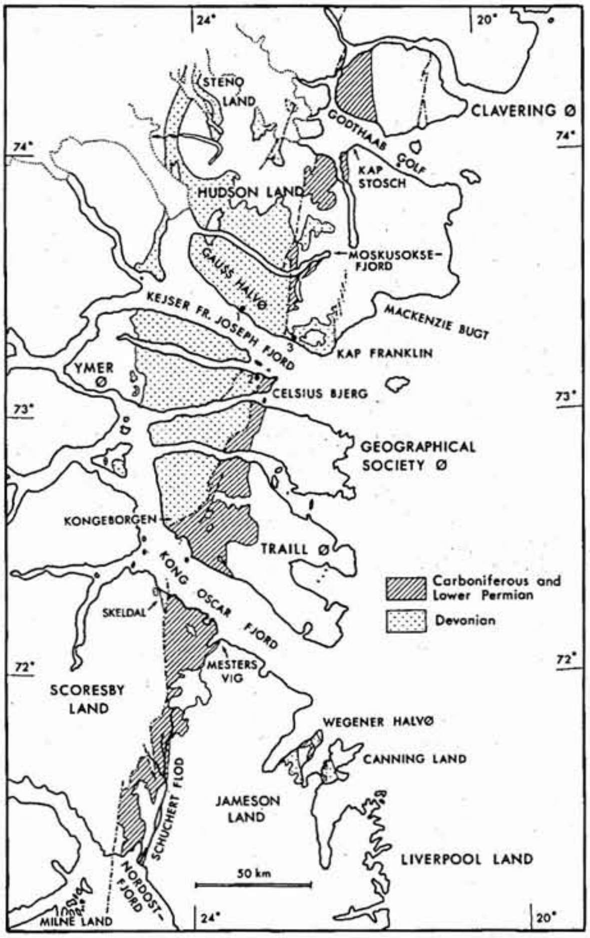

Fig. 1. Continental Devonian deposits in central East Greenland range from the Givetian to the Fammenian-Strunian. Vertebrate fossils, including tetrapods, were collected from deposits of the latter age-bracket, exposed on Stensiö Bjerg and Wiman Bjerg (1) and on Celsius Bjerg (2). At Margrethedal (3) elasmobranchs and actinopterygians were collected from outcrops of the marine Upper Permian. Map from Bendix-Almgreen (1976).

The project was supported from various funds. After flying to Mesters Vig on 5 July the team joined the GGU East Greenland project which provided equipment, provisions and logistic support. On 6 July the team was lifted by helicopter to the first camp site situated at the south-east corner of Stensiö Bjerg. For the subsequent 28 days the team worked from this camp, collecting exclusively from the Upper Devonian deposits cropping out on Stensiö Bjerg and Wiman Bjerg (fig. 1). On 4 August the expedition split into two: one team was transported by helicopter to eastern Ymer Ø where they camped close to the West Circus Valley and the other team proceeded on foot along the north coast of Kejser Franz Joseph Fjord towards Margrethedal, on the east side of which a camp site was chosen for the remaining days of the field season. Both teams were evacuated by helicopter to the GGU Stordal base camp on 16 August.

Upper Devonian vertebrates

The Fammenian-Strunian continental deposits of central East Greenland have yielded a rich fossil vertebrate fauna which in addition to representatives of the cladodontids, acanthodians, arthrodires, antiarchs, dipnoans, porolepiformes, osteolepiformes and struniiformes, also comprises the much celebrated ichthyostegid stegocephalians which until recently were the only known tetrapods of undisputed Devonian age (see e.g. Bendix-Almgreen, 1976; Jarvik, 1980; Clack, in press).

In the summers of 1968–70 expeditions from Cambridge University, under the leadership of Dr Peter Friend, carried out extensive sedimentological field studies of the Devonian deposits of central East Greenland (Friend *et al.*, 1976a,b; Alexander-Marrack & Friend, 1976; Nicholson & Friend, 1976; Yeats & Friend, 1978; Friend *et al.*, 1983). During the field work J. Nicholson succeeded in collecting various tetrapod fossils among which several more or less fragmentary skulls and some postcranial bones came from scree located about 800 m up and close to the south-eastern ridge of Stensiö Bjerg (fig. 2). The material was recently investigated by one of the present writers (JAC) who found that three skulls, preserved juxtaposed to each other in one block of very fine-grained sandstone, all belonged to *Acanthostega gunnari* which had been erected by Jarvik in 1952 (Clack, in press).

Until this discovery *Acanthostega gunnari* was known only from a couple of incomplete skull roofs and the additional material greatly extended the knowledge of the species' skull morphology. The fact that the material from Nicholson's locality also contained other tetrapod remains indicated that this locality had great potential for yielding significant collections of these fossils. Accordingly, the locality was revisited for further collecting during the 1987 field season when not only the rich scree locality was refound, but also the actual horizon of fossiliferous deposits that had yielded the fossils in the scree was located. The deposits belong to the upper part of the Britta Dal Formation (Friend *et al.*, 1983) which in general corresponds to what has earlier been referred to as the Upper Division of the *Remigolepis* series (Jarvik, 1961; Bendix-Almgreen, 1976).

Sedimentology. The Britta Dal Formation (Friend *et al.*, 1983) in which the fossils were found is composed of

Fig. 2. Outcrop of the beds from which tetrapod fossils were collected *in situ* at the south-east crest of Stensiö Bjerg. The sequence, consisting of about 1 m thick channel fill deposits partially laid down on point bars, indicates that deposition took place in shallow meandering channels which apparently formed part of a highly ephemeral low gradient stream system. Tetrapod remains predominate in both number and preservation in the fossil assemblage which also includes plant macrofossils, disarticulated skeletal elements of dipnoans and *Holoptychius* but only a single plate of *Remigolepis*. Photo: H.O. 21.7.1987.

alternating siltstone and very fine to fine-grained sandstone. However, detailed sedimentological investigations are not possible due to pronounced weathering and scree development. The siltstone occupies *c.* 80% and is mainly reddish with some greyish beds. In the upper half of the formation greenish beds also occur. Cross-lamination is the dominant primary structure. Desiccation cracks are common and bioturbation occurs in several intervals. The siltstone is usually highly brecciated either due to frequent desiccation or disruption by rootlets. The sandstone mainly occurs as channel-shaped bodies, 0.5–1 m thick. Individual sandstone-bodies are commonly multistorey. The storeys very often exhibit lateral accretion bedding and fining-upwards lithology. Parallel lamination, cross-lamination and trough cross-bedding dominate the sandstone. Internal mud partings are common, sometimes associated with desiccation cracks. The sandstone was deposited as point bars in shallow meandering fluvial channels. The discharge was highly fluctuating to ephemeral. The depositional environment is accordingly an (ephemeral) meandering channel belt with extensive muddy flood basins.

Tetrapods. Most of the new tetrapod fossils were collected on Stensiö Bjerg from the point bar deposits, forming the outcrop at the 772 m locality and yielding the scree material just below. In addition to a large variety of detached skull and postcranial bones, these comprise a number of articulated skulls of varying degrees of completeness, preservation and size. Some could be identified immediately as belonging to *Acanthostega* and probably pertain to the species *A. gunnari*. However, the bulk of the skull material, which may even comprise *Ichthyostega* specimens, awaits determination. It seems that some of the specimens, preserving skull remains in association with postcranial skeleton parts (including elements of the vertebral column), actually pertain to the genus *Acanthostega*. A specimen showing a sizeable, articulated portion of a body squamation preserved in association with pelvic girdle elements may also belong to this genus (fig. 3). A similar partially preserved body squamation was collected from a locality on Wiman Bjerg where a much weathered and entirely flattened skull of *Acanthostega* (skull length *c.* 9 cm) (fig. 4) was also found preserved in association with articulated parts of the shoulder girdle.

New specimens of the genus *Ichthyostega* were collected from scree derived from the upper part of the Aina Dal Formation (Friend *et al.*, 1983) as exposed just west of Aina Dal. These include part of a skull and associated lower jaw and parts of the pelvic girdle, hind limb (fig. 5) and tail. Yet another partially preserved, but so far unidentified tetrapod specimen showing remains of skull and shoulder girdle in association, was found at the same scree locality. This specimen may derive either from the Wiman Bjerg Formation's lowermost part or perhaps from the Aina Dal Formation.

Ichthyostega skull parts and jaws and other tetrapod remains were also collected from localities on the north side of Celsius Bjerg on eastern Ymer Ø.

Osteolepiformes, Porolepiformes and Stuniiformes. New specimens of the large osteolepiform *Eusthenodon*

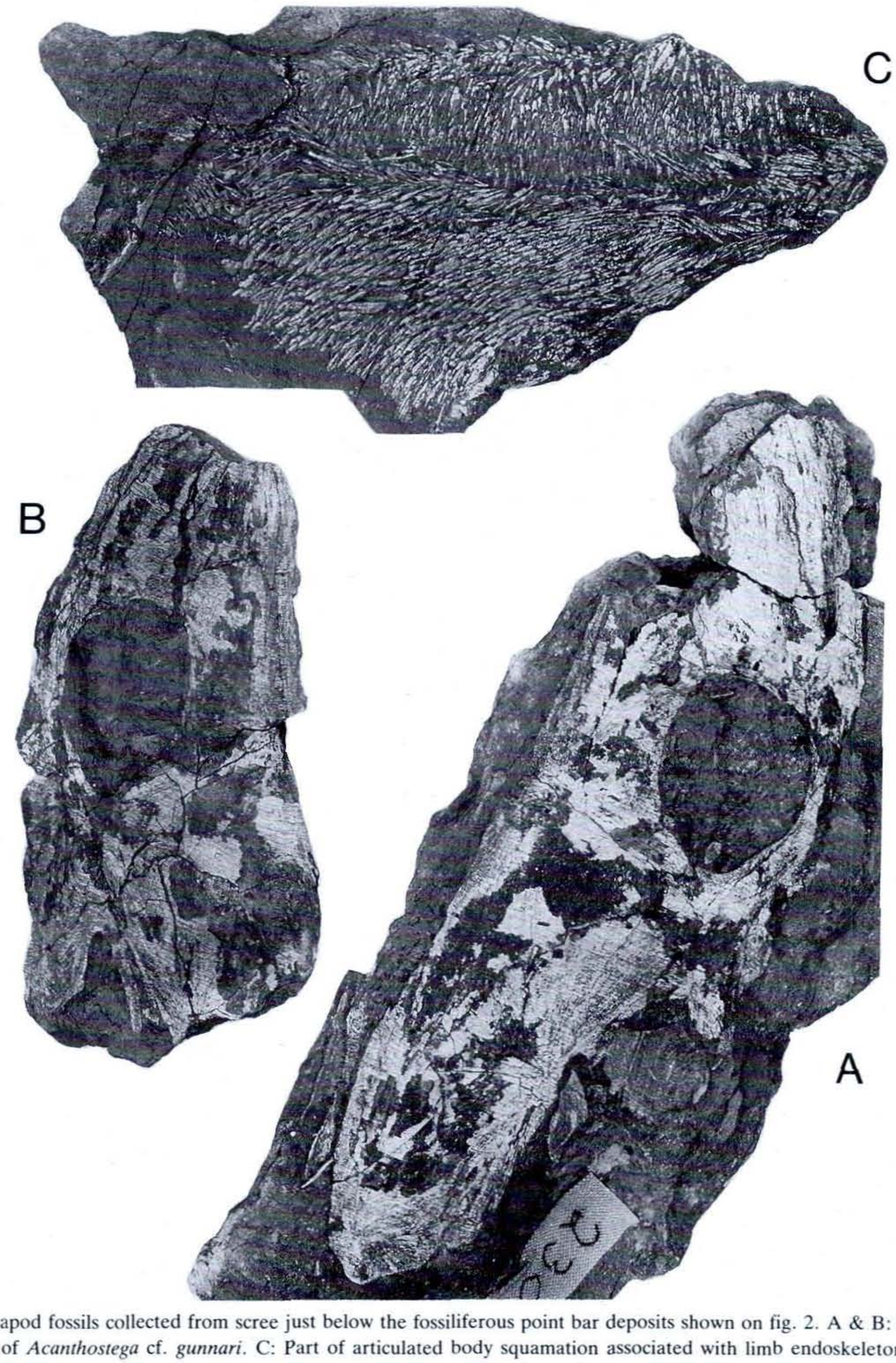

Fig. 3. Tetrapod fossils collected from scree just below the fossiliferous point bar deposits shown on fig. 2. A & B: Incomplete skull roofs of *Acanthostega* cf. *gunnari*. C: Part of articulated body squamation associated with limb endoskeleton elements. Specimens shown in approx. nat. size. A-C: Geological Museum Copenhagen field nos 230a, 251 and 252a, respectively.

Fig. 4. *Acanthostega* sp.; flattened, much weathered skull roof of a small individual collected on Wiman Bjerg. Approx. nat. size. Geological Museum Copenhagen field no. 1400a-b.

Fig. 5. *Ichthyostega* sp.; partially exposed articulated hind-limb and foot endoskeleton. Aina Dal Formation, Stensiö Bjerg. Approx. nat. size. Geological Museum Copenhagen field no. 1384a[9].

(Jarvik, 1952; Bendix-Almgreen, 1976) were collected at localities both on Stensiö Bjerg and on the north side of Celsius Bjerg. Most of the collected rhipidistian material pertains, however, to the porolepiform genus *Holoptychius* (Jarvik, 1972) and includes specimens displaying skull roof, cheek and lower jaw bones, which may provide new information concerning the entire skull bone pattern. Other specimens may give new information regarding the fin endoskeleton. The material derives from localities on Stensiö Bjerg and on the north side of Celsius Bjerg. One locality discovered on Stensiö Bjerg showed an *in situ* concentration of a considerable number of *Holoptychius* individuals preserved articulated, but totally compacted.

A single *Onychodus* parasymphysial tooth whorl deserves mention because its occurrence on Stensiö Bjerg, in scree derived either from the top of the Wiman Bjerg Formation or from the Britta Dal Formation above, establishes the presence of this struniiform genus at higher stratigraphical levels in the East Greenland Upper Devonian than hitherto recorded (Stensiö, 1936).

Dipnoi. New skull material of the genus *Soederberghia* (Lehman, 1959; see also Bendix-Almgreen, 1976) was collected. Other specimens show a partial skull roof, and the palate, ribs, vertebral centra, anal fin and other associated parts possibly from the same genus. This material from localities on the north side of Celsius Bjerg supplements that found on Stensiö Bjerg which probably derives from deposits of the Aina Dal Formation and includes specimens showing parts of vertebral columns. Specimens from deposits of the Britta Dal Formation comprise detached skull roof bones and parts of the palate; among them a pterygoid and associated tooth plate possibly representing the genus *Oervigia* (Lehman, 1959). Significant dipnoan remains also occur in the richly fossiliferous point bar deposits which yielded a large part of the considered tetrapod fossils.

Arthrodires and Antiarchs. A few exoskeletal bones of unidentified arthrodires were collected from scree on Stensiö Bjerg (? Britta Dal Formation) and an arthrodire jaw was found in scree on the north side of Celsius Bjerg. A few specimens of the very common antiarch *Remigolepis* (Stensiö, 1931; Jarvik, 1985, p. 4) preserved articulated, were taken from scree localities on both Stensiö Bjerg and on the north side of Celsius Bjerg.

Plant fossils and palynological samples. Some stem sections derived from the point bar deposits of the 772 m locality on Stensiö Bjerg were considered worth laboratory inspection but it has turned out that only one of these might be identifiable (Dianne Edwards, personal communication 1987). Samples for palynological inspection were also brought back from this locality but processing in the laboratory gave no identifiable spores (Stefan Piasecki, personal communication 1987).

Notes on tetrapods and habitat. The fact that the large number of *in situ* tetrapod fossils from the point bar deposits exposed at the 772 m locality on Stensiö Bjerg comprises fairly complete crania and larger associated parts of skeletons indicates that the tetrapod carcasses had not been subject to any long transport before they became embedded in the deposits. Accordingly it seems reasonable to assume that the deposition site was located within the very area in which the tetrapods lived. This area was characterised by shallow, meandering fluvial channels.

The tetrapods may have searched for their prey mainly in the meandering and intertwining streams where mating and breeding also took place. However, the habitat for these early tetrapods may have included the land immediately around the stream banks. Direct evidence for this in the form of fossil trackways like those known from the Upper Devonian of Australia (Warren & Wakefield, 1972) has not yet been found in East Greenland. However, the structure of the postcranial endoskeleton of *Ichthyostega* indicates that this had the functional prerequisites for body support and locomotion also out of the water and this may have been the case also with *Acanthostega*. There are features suggesting that *Ichthyostega* physiologically was dependent on cutaneous respiration (Jarvik, 1980, p. 226) and *Acanthostega* may have had similar requirements. It should be noted therefore, from finds of fossil plants including impressions of large-size trunks, and from other evidence such as brecciation of siltstone due to disruption by rootlets, that a fairly diversified flora existed in the area during the late Devonian. This can only mean that on the moist ground adjacent to the streams there was enough vegetation to give the early tetrapods which ventured out of the water, adequate shelter also against the desiccating effect of the wind upon their skin.

The habitats and mode of life of such early tetrapods as *Acanthostega* and *Ichthyostega* may to some extent have differed from each other though hardly radically so. Bjerring (1985, p. 44) suggests that "*Ichthyostega*... could hardly walk", since in his interpretation of its pelvic appendicular endoskeleton, *Ichthyostega* lacks the articulatio cruropedalis (see also Jarvik, 1980, figs 163–164). This may be so, but the suggestion that *Ichthyostega* could not walk seems to disagree with almost all the structural features characteristic of the postcranial endoskeleton of this genus. According to Jarvik

(1980), the entire thoracal, abdominal and sacral portion of the vertebral column forms an arched and probably fairly rigid structure apparently capable of supporting the body. The articulation between humerus, ulna and radius is generally like that in other early tetrapods. There are well developed articular connections between sacral vertebrae and the pelvic girdle. This latter forms a large supporting structure and the shape of its two halves suggests that the attaching leg musculature was strongly developed. The compact femur abuts proximally against the well developed acetabulum and has distally articular areas which, judging from their position and shape, suggest that the articulatio genus formed an angle as in other early tetrapods generally considered capable of walking.

Although the articulatio cruropedalis may be lacking in *Ichthyostega*, the shape and other features of the articulations between the middle and distal tarsals, and between the distal tarsals and the metatarsals suggest that the hind feet with their muscle-clad, skin-covered metatarsals and phalanges resting on the ground were fully capable of walking when this tetrapod emerged on land.

Upper Permian marine vertebrates

Collections were made from Upper Permian marine deposits (the Foldvik Creek Formation: Maync, 1942; Birkelund & Perch-Nielsen, 1976) exposed in the deep ravines located on Vestreplateau and Østreplateau on both sides of Margrethedal. Fossil fish remains from the deposits in this area have been reported previously by Maync (1942) and some few palaeoniscoid specimens were collected by Bütler in the late 1940s and early 1950s.

The new fish fossils were, like those just referred to, collected exclusively from the *Posidonia* Shale Member (Birkelund & Perch-Nielsen, 1976). They occur here both in the shale proper and in the often very large concretions forming horizontal bands in the shale. The shale locally attains a considerable thickness on Østreplateau, as shown by the exposures around and between rivers F and G (for geographical details see Bütler, 1954, Pl. 6). In general the occurrence and preservation of the vertebrate fossils correspond closely to those encountered in the Kap Stosch area towards the north (Nielsen, 1935; Bendix-Almgreen, 1976) and, as a whole, the fossil content of the *Posidonia* Shale and its concretions is also similar.

The bivalve *Posidonia permica* frequently occurs in large concentrations on slabs of the shale in which there are also concentrations of cephaloped prehensile arm-hooks representing no doubt distorted natural assemblages from single individuals. A specimen of *Macrotheca almgreeni* (Peel & Yochelson, 1984) was also found. This hyolithid mollusc has hitherto been reported only from the Kap Stosch exposures where it also occurs in the *Productus* Limestone Member of the Foldvik Creek Formation which here is intercalated with the *Martinia* Limestone Member (Maync, 1942, fig. 17 & Pl. 6; Birkelund & Perch-Nielsen, 1976).

Elasmobranchii. Among these the edestid *Fadenia crenulata* is represented by specimens showing articulated large parts of skulls and the associated dentition, and a single specimen preserving a part of an articulated caudal fin endoskeleton and the associated scale covering. Another of the edestids, *Erikodus groenlandicus*, is represented by large parts of a skull and the dentition. The petalodontid *Janassa*, either the species *J. kochi* or *J. unguicula*, can now also be recorded as part of the elasmobranch fauna of this area as in that of the Kap Stosch area (Bendix-Almgreen, 1976).

Actinopterygii. The palaeoniscoids, represented by both incomplete specimens, detached bones and several virtually complete specimens, include apparently also some juveniles. According to preliminary inspection in the field, the material appears to comprise species referable to the genera *Elonichthys*, *Pygopterus*, *Palaeoniscus*, *Boreolepis* and *Platysomus*.

The new material suggests a faunal composition similar to that of the Kap Stosch area (Bendix-Almgreen, 1976). There is also similarity with the elasmobranch-palaeoniscoid assemblage from localities in Jameson Land and Wegener Halvø, towards the south, known from collections made during GGU's field survey and inspected by one of us (SEB-A).

During the field work in the Margrethedal area several specimens of the actinopterygians *Bobasatrania*, *Pteronisculus* and *Boreosomus*, as well as a single selachian (? *Polyacrodus*), were collected. All these derive from the concretions characteristic of the marine early Triassic (Lower Scythian) deposits which are quite extensively developed within the area.

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Glacier velocities from aerial photographs in North and North-East Greenland

A. K. Higgins

General descriptions of the glaciers of North and North-East Greenland have been given by Koch (1928), Davies & Krinsley (1962) and Weidick (1975). These descriptions, however, provide little in the way of quantitative data on glacier velocities, although Davies & Krinsley concluded that a large number of glaciers and

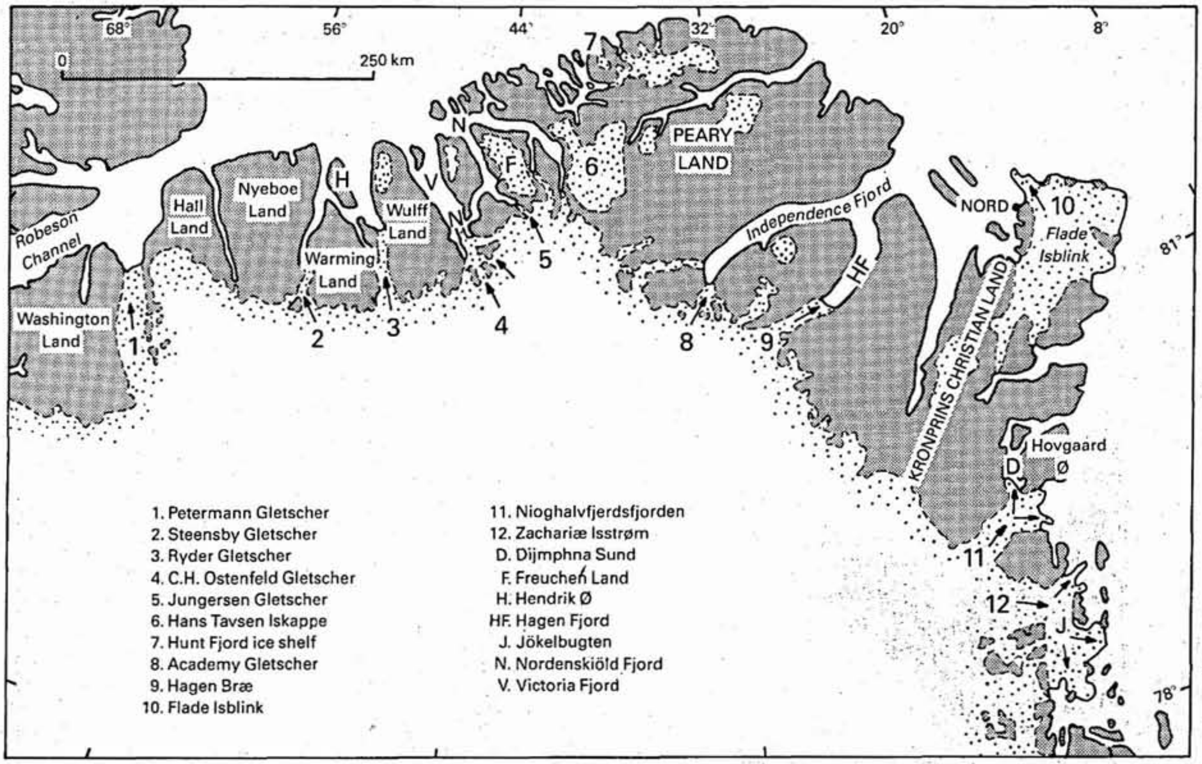

Fig. 1. Map of North and North-East Greenland showing location of glaciers described in text. Land areas are shaded. The marginal area of the Inland Ice and independent ice caps and glaciers have a dotted ornament.

small ice caps in North Greenland exhibited stable conditions, with a significant number showing evidence of recent retreat. Comparisons of vertical aerial photographs taken in 1959–63, 1971 and 1978 permit measurements of glacier velocity to be made on floating ice tongues which have preserved a distinctive surface pattern of meandering streams and crevasses. These show the largest glaciers draining the Inland Ice in North and North-East Greenland to have average velocities ranging from 300 to 900 m/year.

This study of vertical aerial photographs has also demonstrated that for floating glacier tongues the position of the glacier terminus is not a reliable indicator of advance or retreat. In the fjords of North Greenland semi-permanent sea ice often maintains the integrity of advancing floating glacier fronts for periods of ten to twenty years (Koch, 1928; Weidick, 1975); the break-up of the floating tongue in rare summers when the sea ice melts completely may give the impression of a sudden retreat, but this 'retreat' is unrelated to changes in the mass balance.

Petermann Gletscher (fig. 1: no. 1) is 110 km long and 15 km wide at its front, and occupies a deep cleft between Washington Land and Hall Land. At least the northern 40 km of the glacier is afloat, and matching of surface features on aerial photographs from 1959 and 1978 demonstrate that the central part of the glacier has moved 17 km in 19 years, an average velocity of 895 m/year.

The position of the glacier terminus is reported to have had a fairly stable position since 1876 (Koch, 1928; Davies & Krinsley, 1962), which indicates that large icebergs must regularly calve from the front and escape into Robeson Channel, where there is often open water in the summer.

Steensby Gletscher (fig. 1: no. 2) drains into the head of Sankt George Fjord, between Warming Land and Nyeboe Land. It is 62 km long, and about 3.5 km wide at the front where it characteristically breaks up into several floating lobes separated by aggregates of small icebergs (Ahnert, 1963). Aerial photographs from 1963, 1971 and 1978 demonstrate a velocity of 430 m/year. Ahnert's estimate of an annual advance of 5.4 to 8.7 km was based on an incorrect assumption. The semi-permanent sea ice has maintained the advancing front of Steensby Gletscher as an intact mass, and no large icebergs drifted away from the front between 1953 and 1978.

Ryder Gletscher (fig. 1: no. 3) is more than 80 km long and drains northwards between Wulff Land and Permin Land, where it is 8 km wide. Characteristic surface features on the frontal floating segment of the glacier demonstrate a movement of up to 8 km between 1963 and 1978, an average velocity of about 535 m/year.

Lauge Koch's observations in 1917 showed the floating glacier tongue to extend a considerable distance farther north than at present, into the fjord between Hendrik Ø and Wulff Land (Koch, 1928). Davies & Krinsley (1962) note that this tongue had broken up prior to 1947, and that the position of the glacier front withdrew a further 5 km from 1947 to 1956. Between 1971 and 1978 the front has advanced without drift of significant icebergs away from the front.

Break-up and dispersal of the floating glacier tongue clearly takes place only in rare summers when the sea ice melts completely. Oblique aerial photographs from 1953 show a collection of large icebergs off the front of Ryder Gletscher, which in an ice-free summer prior to 1963 had drifted 15 to 45 km northwards to positions between Hendrik Ø and Wulff Land. These tabular icebergs occupied virtually unchanged positions on 1963 and 1978 aerial photographs and in 1985 (personal observations), showing the sea ice had not melted completely for at least 22 years.

C. H. Ostenfeld Gletscher (fig. 1: no. 4). In May 1917 Lauge Koch encountered in the outer part of Victoria Fjord what he initially took to be glacier ice from the floating tongue of C. H. Ostenfeld Gletscher, but later concluded it might be old sea ice (Koch, 1928). His first impression was correct, and as shown by Davies & Krinsley (1962) the inner 75 km of Victoria Fjord is tightly packed by large and small icebergs derived from the glaciers at the head of the fjord. The position of the iceberg front is currently much the same as in 1917.

Six large glaciers merge at the head of Victoria Fjord, of which the most important is C. H. Ostenfeld Gletscher; it is up to 7 km wide and has a more or less connected floating segment projecting for 25 km into Victoria Fjord. Velocity measurements made on the basis of aerial photographs from 1963, 1971 and 1978 range from 750 to 815 m/year. Two glaciers west of C. H. Ostenfeld Gletscher are inactive, whereas the three glaciers to the east have velocities of 400–550 m/year.

Jungersen Gletscher (fig. 1: no. 5). This is not a large glacier filling Nordenskiöld Fjord as depicted by Koch (1928), but as shown on modern maps it is a relatively small glacier 2.5 km wide draining from the Inland Ice south of Freuchen Land into the head of the fjord. The measured velocity determined from 1963 and 1978 vertical aerial photographs is 350 m/year.

Hans Tavsen Iskappe (fig. 1: no. 6) is an independent ice cap about 70 km across. Three small outlet glaciers draining into the fjord to the west have measured velocities of 40–70 m/year, and a larger glacier draining northwards 100 m/year. A drilling operation was carried out on Hans Tavsen Iskappe during the 1975 Greenland Ice Sheet Programme (Langway *et al.*, 1985).

Hunt Fjord ice shelf (fig. 1: no. 7). Small ice shelves occur in northern Peary Land, one of which completely fills Hunt Fjord; a smaller ice shelf partially blocks the sound west of Hunt Fjord. These are the only ice shelves in North Greenland which bear comparison with the better known ice shelves fringing northern Ellesmere Island (Jeffries, 1987). Like them they appear to be very ancient features and exhibit a comparable surface pattern of undulating ridges and water-filled hollows. However, the North Greenland ice shelves are clearly fed by the alpine glaciers in north Peary Land, and while they generally show little sign of movement between 1963 and 1978 aerial photographs, a glacier feeding a portion of the west front of Hunt Fjord ice shelf is moving forward at 23 m/year.

Academy Gletscher (fig. 1: no. 8). Early observations of the 10 km wide glacier at the head of Independence Fjord indicate a floating frontal portion of hummocky ice and densely packed icebergs, extending about 12 km into the fjord (Peary, 1898; Freuchen, 1915; Koch, 1928). This floating portion of Academy Gletscher had dispersed by 1956 (Davies & Krinsley, 1962), and aerial photographs from 1962 and 1978 show no indication of re-establishment. It is inferred that since the 1950s the fjord ice has melted sufficiently frequently that floating segments of the glacier disintegrate and disperse before they can form substantial ice tongues. No velocity measurements are possible on aerial photographs of the present crevassed frontal portion.

Hagen Bræ (fig. 1: no. 9), at the head of Hagen Fjord, is 75 km long and about 10 km wide in its central part. Towards the front the glacier widens, and while the southern part is partially dammed by two islands and breaks up, the northern part of the front continues into the fjord as a floating segment 6 km wide and up to 18 km long. Davies & Krinsley (1962) describe the floating glacier tongue as stagnant with a surface of interlacing streams and large interconnecting ponds in parallel troughs. Aerial photographs from 1960 and 1978 show that Academy Gletscher with its floating front is moving outwards at 540 m/year. Large tabular icebergs up to 5 km by 2 km in size, which formed the front of the glacier tongue in 1960, had broken free and drifted up to 45 km towards the outer part of Hagen Fjord by 1978.

Flade Isblink (fig. 1: no. 10). Northern Kronprins Christian Land supports a large independent ice cap, Flade Isblink. It is more than 100 km long and up to 75 km wide but has only a few outlet glaciers which reach the sea, the most important being that east of Station Nord where a floating glacier up to 25 km broad extends northwards into the sea for up to 15 km. The eastern lobe of this floating glacier moves at 175 m annually and the western lobe at 360 m annually. There was no significant loss from the advancing front between 1961 and 1978, although extensive open water leads were adjacent to the front on the 1978 aerial photographs.

Nioghalvfjerdsfjorden (fig. 1: no. 11). The interior of Nioghalvfjerdsfjorden is filled by an extensive floating glacier tongue, of which a northern branch projects into Dijmphna Sund west of Hovgaard Ø. The main glacier is 60 km long and 18 km wide at its narrowest point. It widens eastwards and the outer 30 km segment is afloat. A prominent system of undulating ridges and hollows, first described by Koch & Wegener (1911), is developed on the surface of the floating glacier.

The velocity of parts of the main glacier have been measured from 1962 and 1978 aerial photographs at 310–330 m/year. The main floating front of the glacier in Nioghalvfjerdsfjorden abuts against a series of small islands. On the north side of the fjord movements of only 35–40 m/year have been measured, whereas on the south side of the fjord tongues of ice projecting between islands move seawards at rates of up to 160 m/year. The branch of the glacier moving north into Dijmphna Sund has an average velocity of 210–230 m/year.

Zachariæ Isstrøm/Jökelbugten (fig. 1: no. 12). The largest area of floating glacier ice in Greenland is that filling Jökelbugten, which emanates from Zachariæ Isstrøm. This area of ice-covered sea was initially described by Koch & Wegener (1911) as 'Das schwimmende Indlandeis der Jökelbugt', and was classified by Koch (1928) as 'confluent ice'. The floating glacier ice cover a region 100 km from north to south and a maximum of 50 km from east to west.

Zachariæ Isstrøm has a minimum width of 20 km, and the central active stream has a velocity estimated from 1963 and 1978 vertical aerial photographs to be at least 470 m/year. The northern part of the floating glacier ice is a single intact mass whose outward expansion is hindered by large and small islands. Between the islands narrow tongues of shelf ice up to 15 km long move outwards at 220–280 m/year. The southern area of floating glacier ice filling Jökelbugten comprises tightly packed, broken, tabular icebergs. Movement of individual icebergs varies from 66 m to 600 m/year averaged over 15 years, the rate of movement being clearly related to the damming effect of islands, and large icebergs trapped between islands.

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Surging glaciers in Greenland – a status

Anker Weidick

New wide-angle vertical aerial photography covering most of West and East Greenland were flown in the years 1981 and 1985 by Mark Hurd Corp., Minneapolis, Minnesota, for the Geodetic Institute, Copenhagen. The photographs are on a scale of approximately 1:150 000, so that single photographs cover a large area, and are valuable as bench marks for glacier changes. These photographs have been used for updating information on West Greenland glacier changes and the history of West Greenland glacier surges. Those covering East Greenland have been used for location of important centres of surging glaciers.

Procedure

Glacier surges are defined as a mass "transfer of a large volume of ice from a reservoir area to a receiving area" (Paterson, 1981, p. 279). Usually surges are located by diagnostic features resulting from the surging activity, such as strong crevassing, complex lobation of the frontal areas, or widespread pitting of the glacier surface. Since these features 'heal' with time a full coverage of glacier surges in a given region requires repeated aerial photography over several decades. This has been possible in West Greenland where the glacier surges concentrated around Disko island and Nûgssuaq peninsula are well documented.

In East Greenland new photographic coverage in 1981 has pinpointed areas around the Blosseville Kyst as having the highest frequency of glacier surges in Greenland. The possibility of unravelling the surge history is more restricted here and has not yet been attempted; the present article indicates some of the trends.

Disko and Nûgssuaq

Both areas were among the first to be mapped by the Geodetic Institute (map sheets 1:250 000 in 1931/33), and since 1942 they have been covered by aerial photography on several occasions. About a score of glaciers exhibit more or less pronounced morphological criteria of glacier surges, but at only half of them can surge events be documented and dated.

Individual localities. On the basis of the documented surges and delay in healing of surge features, recessional rates of the glacier fronts during quiescent phases, and geomorphological ice-contact features of the foreland, an approximate age of older surge events can be suggested for some of the glaciers (fig. 1). Nearly all the surging glaciers are located in areas of Tertiary basalt overlying and intercalated by Cretaceous sediments. Only a single glacier (1IB26003–Agssakait sermia) is situated in the area of Archaean gneiss.

In fig. 1 the position of the glacier fronts are generally shown as the distance from the neoglacial outer moraines. Exceptions are 1HD06030–Stordal and 1HB15012–Kuanerssuit where the reference points are at a supposed minimum of glacier advance in the last 100–200 years but where maximum extension of the neoglacial advances might be essentially greater. The frontal advances are shown with arrows. Hatched arrows indicate documented and dated advances, and plain arrows those suggested from the evidence described above. The extent of the advances is indicated by the position of the arrow point.

The frontal recession in the quiescent phase is at rates of 10–100 m per year, but thinning of the glacier is usually more pronounced. This implies that the outer glacier lobe of some of the glaciers can become detached over thresholds and then transformed into dead ice. During the surges, the reactivated glacier lobes in some cases move over old glacier ice, and in one example (1HE09051–Kûganguaq head) the advancing glacier front does not extend beyond the margin of the ice of the former glacier, so that two contemporaneous, recessional curves have to be drawn.

General trends of the surges. Information on the eleven surging glaciers is mainly restricted to the most recent surge events, but even here the evidence for the events of the glaciers on Agatfjeldet (1HE09090 and 1HE09095) is questionable. For three glaciers (1HB10013, 1HD06030 and 1IA02034) two events, and for one glacier (1HB10036, cf. fig. 2) three events can be suggested. The quiescent phase for all the examples given seems to be around thirty to fifty years while the surge event lasts for up to a few years.

The magnitude of the frontal variations due to the surges vary between one and a little more than two kilometres. A general feature is that the surges in the last century did not reach the neoglacial maximum and that there is a tendency for subsequent surge advances

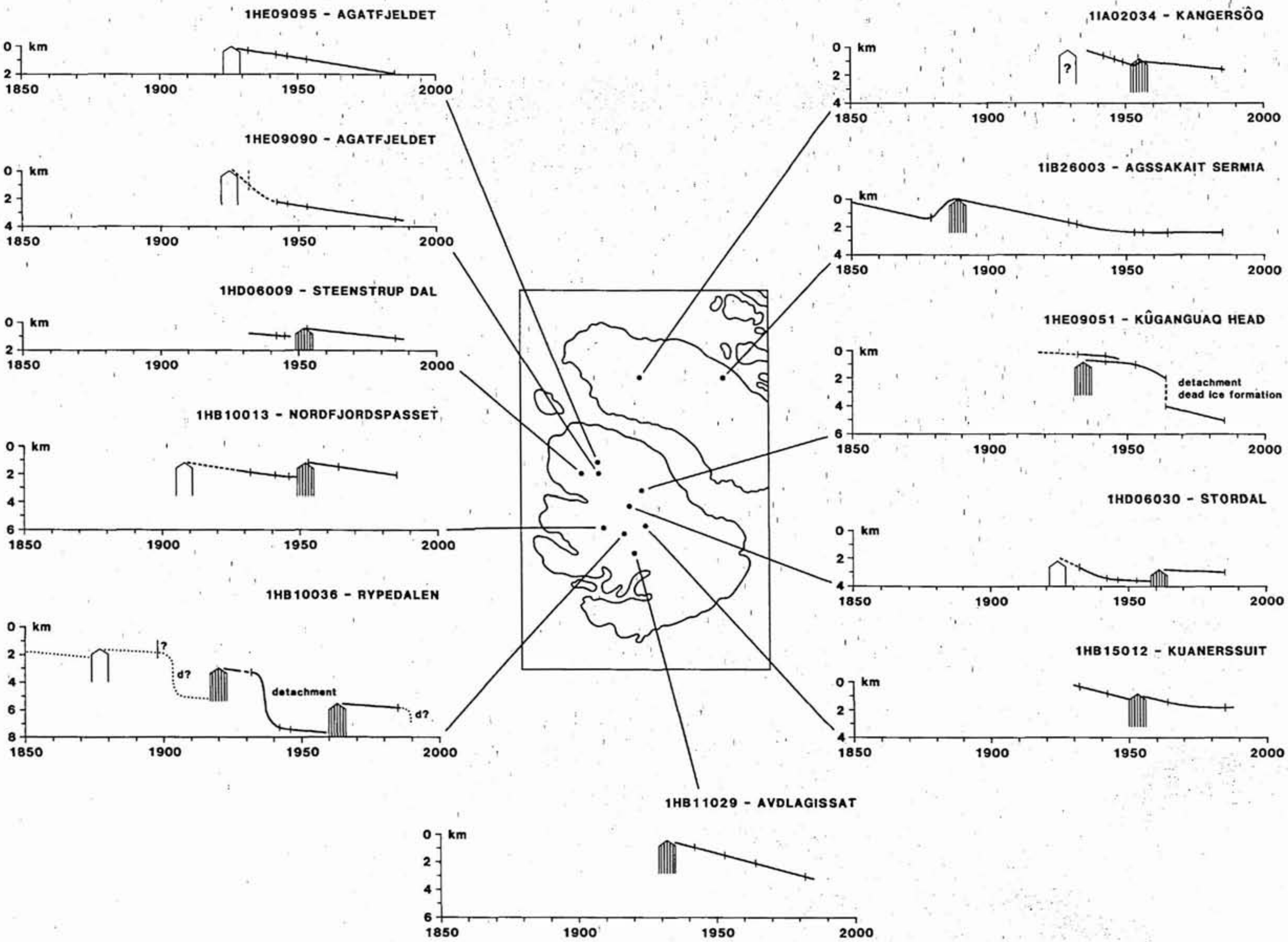

Fig. 1. Frontal fluctuations of surging glaciers on Disko and Nûgssuaq, West Greenland.

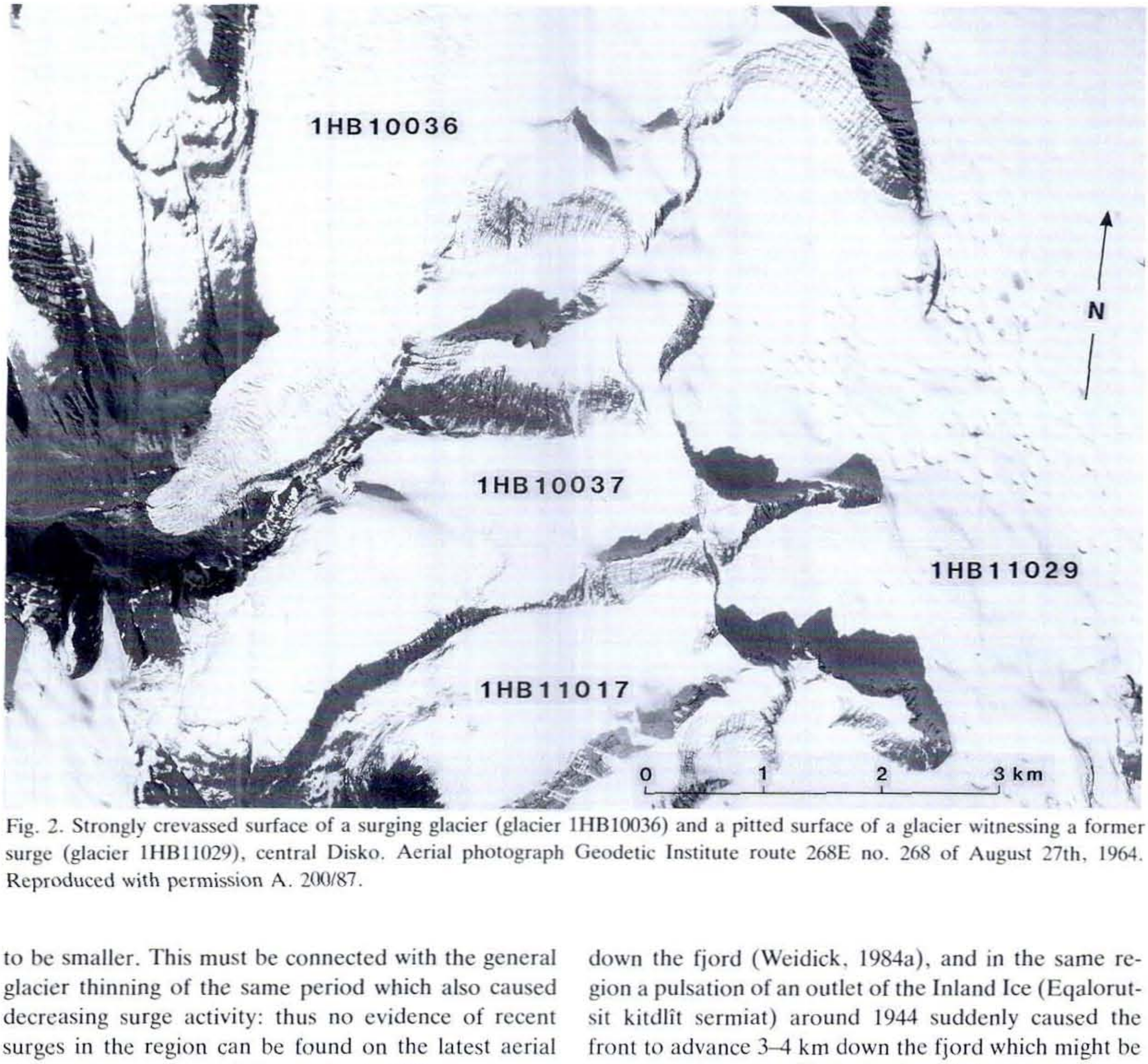

Fig. 2. Strongly crevassed surface of a surging glacier (glacier 1HB10036) and a pitted surface of a glacier witnessing a former surge (glacier 1HB11029), central Disko. Aerial photograph Geodetic Institute route 268E no. 268 of August 27th, 1964. Reproduced with permission A. 200/87.

to be smaller. This must be connected with the general glacier thinning of the same period which also caused decreasing surge activity: thus no evidence of recent surges in the region can be found on the latest aerial photograph coverage of the region in 1985.

Evidence which could have been interpreted as relict structures has been found on a number of glaciers other than those described. Possibly many glaciers of the region passed through a period of optimum conditions for surging during the 'Little Ice Age' and the observations presented here just indicate the end of this period for the glaciers in the region.

Other parts of West Greenland

In spite of extensive work on the registration of all West Greenland glaciers, little evidence of surging has been found outside the regions of Disko and Nûgssuaq. Surge behaviour in the middle of the 19th century may possibly have brought the outlet of an ice cap, Sermeq in Søndre Sermilik fjord in South Greenland, 12 km down the fjord (Weidick, 1984a), and in the same region a pulsation of an outlet of the Inland Ice (Eqalorutsit kitdlît sermiat) around 1944 suddenly caused the front to advance 3–4 km down the fjord which might be interpreted as surge-like behaviour (Weidick, 1984b).

East Greenland

The first observation of surges in East Greenland was made during GGU field work in the Scoresby Sund region, where descriptions of the glacier Løberen were given by Henriksen & Watt (1968) and Olesen & Reeh (1969), and of Bjørnbo Gletscher by Rutishauser (1971). Both glaciers are situated in the Stauning Alper, a mountainous area of Caledonian migmatites and granites (Henriksen & Higgins, 1976, especially map fig. 198, p. 218). The length of the quiescent phase has not been determined in the case of Løberen where the documented surge took place in the early 1960s, but a preceding quiescent phase must have covered several decades or even a century of downwasting of a 7 km

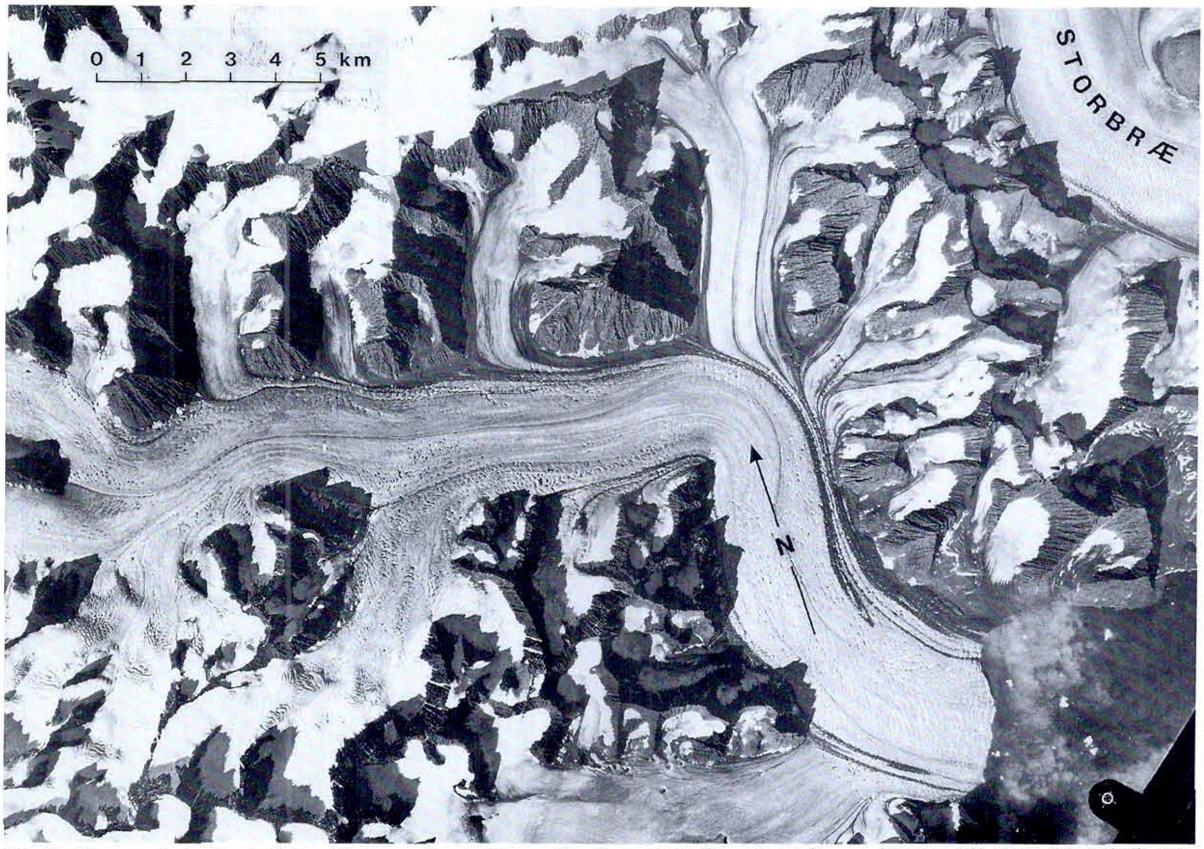

Fig. 3. Unnamed glacier at Johan Petersen Bugt, Blosseville Kyst, East Greenland (approximately 68° 50′N, 26° 30′W). The surface is strongly pitted after a former surge. Aerial photograph, Geodetic Institute route 878 B no. 1964 of August 14th, 1981. Reproduced with permission A. 200/87.

long glacier lobe with a maximum thickness of at least 300 m. As far as Bjørnbo Gletscher is concerned, Rutishauser (1971, p. 236) suggests an interval of 100 years between a surge around 1890 and an expected one at around 1990.

More surges can probably be documented in the Scoresby Sund region, since a number of glaciers indicate structures which might be interpreted as results of surges. However, the greatest concentration of active surging glaciers can be observed on aerial photographs of the stretch between Scoresby Sund and Kangerdlugssuaq. Here a study of the most recent aerial photographs from 1981 (fig. 3) gives good evidence (crevassing, pitting and lobation of moraines) of surging events at 26 localities. The coverage of older photographs of the region is probably insufficient for detailed documentation of events, and thus the glaciers plotted in fig. 4 must indicate a minimum of surges in the region. A widespread pitting of the firn area of the large unnamed glacier south of Torvgletscher in this region is described by Rucklidge (1966), but without connecting their origin to surges.

The area of Scoresby Sund – Kangerdlugssuaq is built up of Tertiary basalts as in the case of the Disko–Nûgssuaq area on the west coast.

Regional occurrences of surges in Greenland

Although no systematic investigation of surging glaciers has been made in Greenland, the examples described here are evidence of widespread occurrence of this behaviour, especially in the basalt provinces of West and East Greenland. This might imply a connection of many surges to special conditions of permeability or roughness of the subsurface, as pointed out by Post (1969) for glacier surges in North America.

The surging glaciers range from modest cirque glaciers to great outlets of ice fields or ice caps, which in East Greenland form complex systems of merging ice streams.

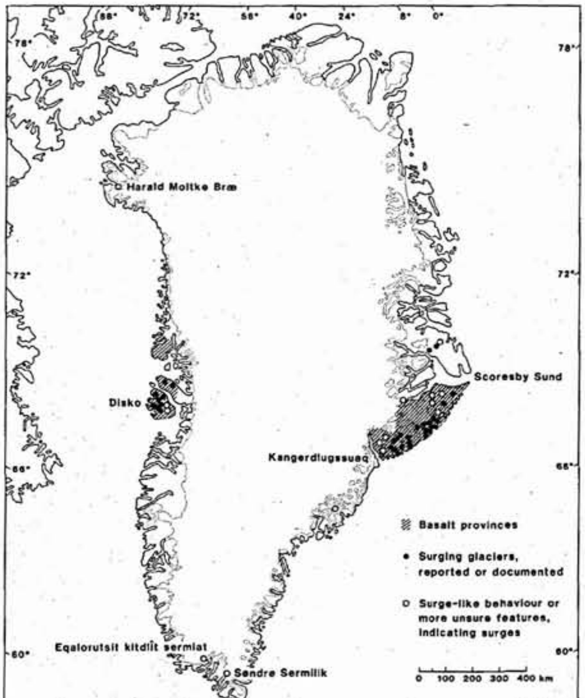

Fig. 4. Surging glaciers in Greenland.

Classification of surges according to time

Dating of events is possible for only a few of the surging glaciers. The ice covers of the wide plateaus of Disko seem very sensitive to climatic fluctuations; rise of the climatic snow line by only about 200 m in the last century meant that large areas of the ice bodies here were converted from accumulation areas to ablation areas.

Apart from the general thinning of the glaciers on Disko and Nûgssuaq, it is possible to study an area where glaciers in most recent time pass from a surging phase to a 'normal one'. This is not the case for the East Greenland localities, which are essentially situated in areas where fluctuations of the climatic snow line rarely imply the same consequences as in West Greenland. Specific conditions apply to the Inland Ice margin where only a single event of pulsation of the outlet Eqalorutsit kitdlît sermiat in South Greenland has been documented (Weidick, 1984b).

The ice streams of the Inland Ice have been labelled 'permanent surges', which in a systematic sense can only be connected with the temporary surges described above in so far as there is a gradation between:

(1) Glacier areas with more or less continuous surging (ice streams).
(2) Glaciers with more or less repeated temporary surge events.
(3) Glaciers with a single surge event.

Transitions between (1) and (2) might be found in the East Greenland basalt area of complex systems of outlets from large ice caps, or possibly from the example of the calf-ice producing outlet from the Inland Ice in North-West Greenland (Harald Moltke Bræ). The latter glacier is normally regarded as in permanent surge, but its velocity near the front has varied in this century between 30 m/year and 1 km/year or more (Mock, 1966).

Transitions between (2) and (3) can be either isolated glaciers that only pass through a period of surge conditions for a short time, or where the glacier area is so extensive that a build up for surging takes considerably longer than the 10–100 years (Paterson, 1981, p. 289) usually recorded for a quiescent phase.

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Mass balance, ice velocity and ice temperature at the Inland Ice margin north-east of Jakobshavn, central West Greenland

Henrik Højmark Thomsen

Glaciological field investigations were made on the Inland Ice north-east of Jakobshavn. The work is part of the hydropower investigations at Påkitsoq in a drainage basin proposed for a local hydropower project. Brief reports of the work have been given by Thomsen (1983, 1984, 1985, 1986).

Mass balance measurements 1986/1987

The glaciological programme was started in August 1982 when stakes were drilled into the ice for measuring the mass balance (fig. 1). The stakes were visited by helicopter on 13 May and 13 August. Six of the stakes are located near the ice margin ending in lake 187 (fig. 2) and are represented by stake 2 in fig. 1. The stakes have been plotted on the most recent map of the area, based on aerial photographs from 1985 (see Thomsen, 1988).

The winter snow cover on the ice was very patchy and confined mainly to drifts in gullies and crevasses up to an elevation of about 500 m a.s.l., but snow cover was continuous at higher elevations. The transient balance for the winter period was measured in snow pits and by depth soundings at the stakes. As there were no signs of heavy melting during the winter, the observed distribution of snow cover is probably due to wind drifting. At elevations of about 500 m a.s.l. the 1987 summer ablation is high compared to earlier years.

At lower elevations the stakes were melted out confirming strong melting during the summer. The transient and annual balances are shown in Table 1. In the case of stakes which melted out, a minimum balance figure can be given based on the length of the stake still in the ice at the time of the spring stake readings. These figures are given in brackets in Table 1.

Ice temperature measurements

To measure englacial temperatures two sets of thermistor strings were drilled down to depths of 202 m and 300 m in the ablation area (fig. 2). The holes were drilled with a newly constructed hot water drill (see Olesen & Clausen, 1988). The drilling operation was carried out from a base camp established on the ice. The 202 m hole is situated 4.4 km upstream from the ice margin with an ice-surface elevation of 490 m a.s.l. The

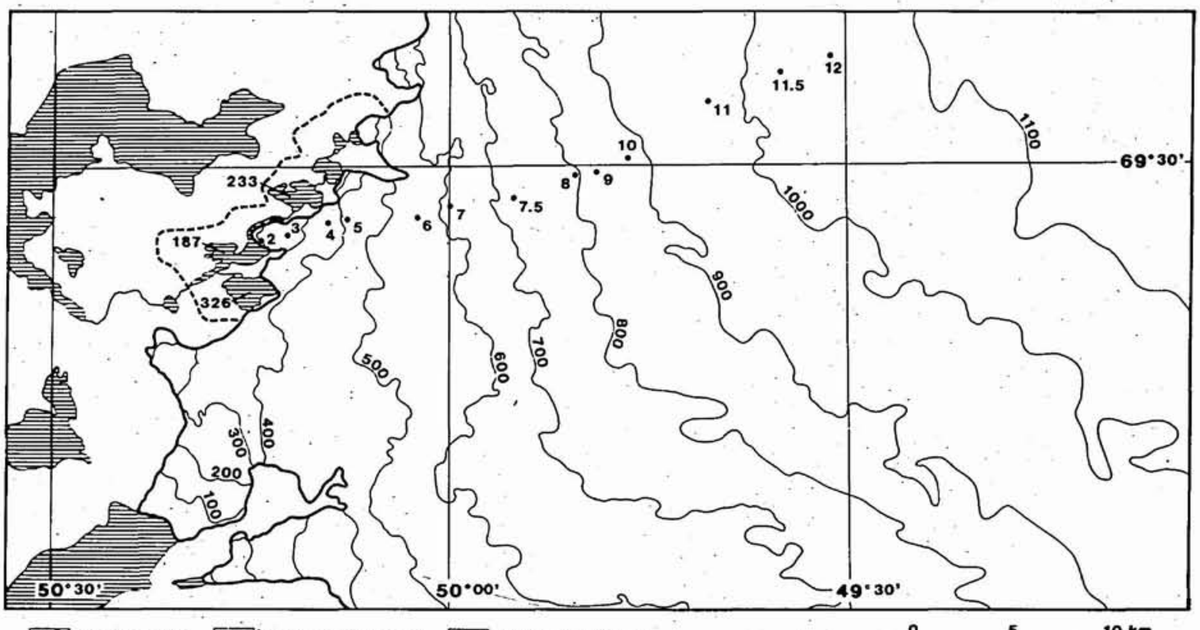

Fig. 1. Drainage basin at Påkitsoq. Stakes drilled into the ice are shown. Contours in metres.

Table 1. Transient and annual balances for the Inland Ice at Pâkitsoq
millimetres of water

Stake	Elevation m a.s.l.	25th Aug 86 13th May 87	13th May 87 13th Aug 87	25th Aug 86 13th Aug 87
2.0	210	−580	>(−2900)	>(−3500)
2.1	205	−850	>(−2900)	>(−3800)
2.2	205	−800	>(−2900)	>(−3700)
2.3	200	−740	>(−2800)	>(−3600)
2.4	200	−810	>(−2800)	>(−3600)
2.5	200	−560	>(−2900)	>(−3500)
3.0	235	−630	>(−2900)	>(−3500)
4.0	380	−570	>(−2700)	>(−3300)
5.0	415	−560	>(−3200)	>(−3800)
6.0	560	−100	−2080	−2180
7.0	615	−180	−1980	−2160
7.5	720	− 70	−2000	−2080
8.0	780	140	−2370	−2230
9.0	850	60	−2270	−2210
10.0	890	130	−1560	−1430
11.0	965	140	−1750	−1620
11.5	1020	140	−1580	−1430
12.0	1070	340	− 480*	− 130*

* estimated.

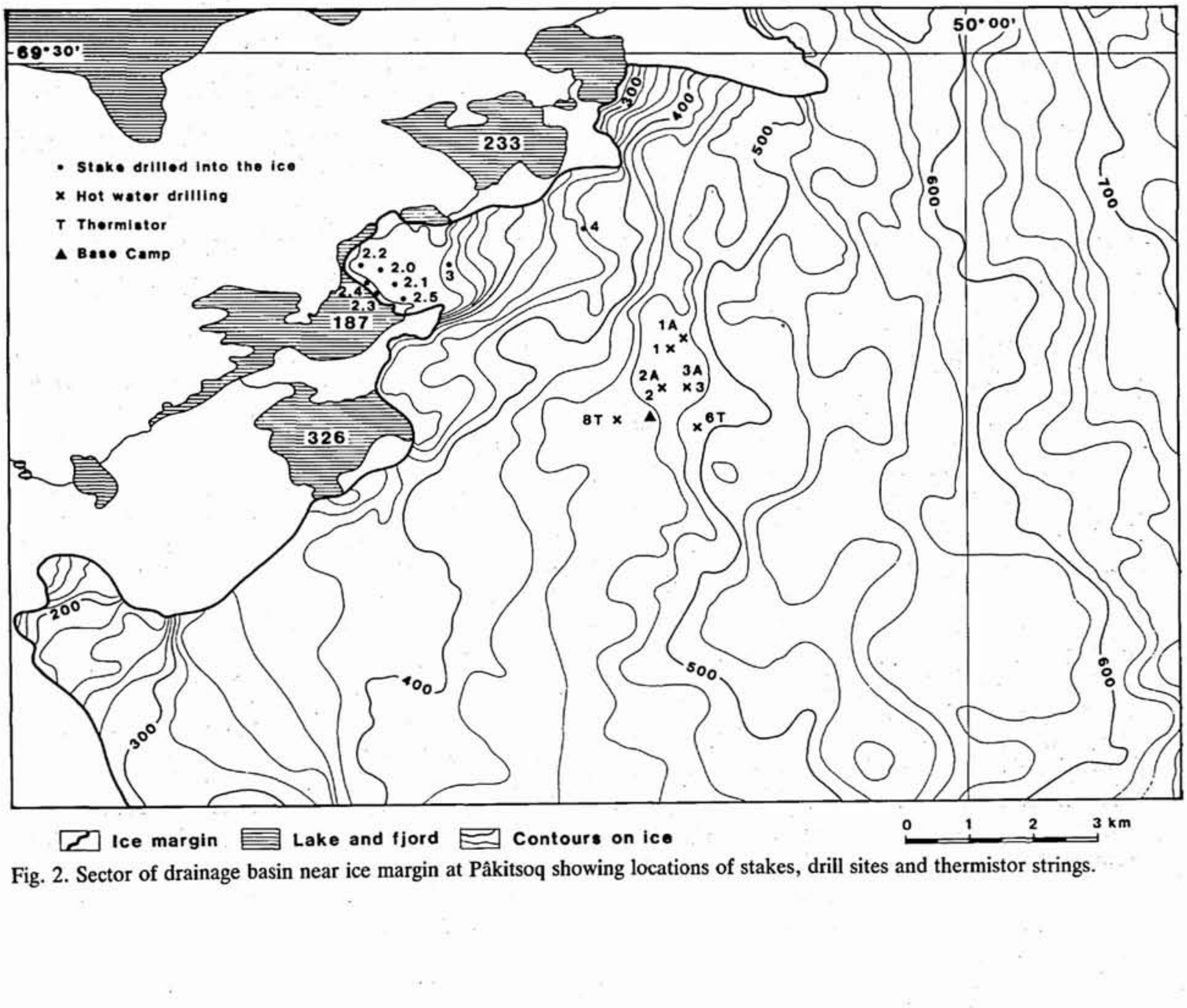

Fig. 2. Sector of drainage basin near ice margin at Pâkitsoq showing locations of stakes, drill sites and thermistor strings.

ice thickness at this location is 300 m according to radio echo soundings (Thorning & Hansen, 1987). The 300 m hole which extends to the bottom of the ice in accordance with radar measurements is situated 3.2 km upstream from the ice margin with an ice surface elevation of 455 m a.s.l.

Thermistors were mounted at every 25 m on the strings except for the lower end where the distance between the three lowest thermistors were 10 and 15 m, respectively. The accuracy of the temperature measurements in the ice is ± 0.2°C. Temperature readings were made several times during the two weeks duration of the drilling operation and were read five weeks later by a visiting field team from Greenland Technical Organisation (GTO). Temperature readings of thermistors drilled into the ice with a hot water drill on White Glacier, Axel Heiberg Island, showed temperatures to be sufficiently close to the equilibrium state to allow measurements after 2 to 3 weeks (Blatter, 1985). The latest readings at Påkitsoq are assumed to be close to the equilibrium state, but exact verification is needed next year. The temperature readings reveal negative temperatures in the whole ice body and a small range through the profiles with a minimum temperature of −2.1°C and a maximum temperature of −0.6°C. Temperature measured at the bottom of the ice is −0.9°C.

Ice velocity measurements

Ice velocity was measured on the glacier tongue ending in lake 187. Ice velocity is measured by theodolite survey at stakes drilled into the ice from fixed points established on the ground. Ice velocities are given in Table 2 and stake locations are given in fig. 2. Ice movement is highest at stake 3 located at the foot of a small icefall. For the stakes near to the ice margin there is generally a higher velocity at the southern lateral part of the tongue compared to the central part. For all stakes there is a marked seasonal variation in ice movement, with mean summer velocities up to twice the mean winter velocity. Variations in sliding velocity could be an explanation (Paterson, 1981; Andreasen, 1985) which in turn implies that the basal ice is at the melting point and that surface meltwater can penetrate to the bed and build up high water pressure. It is not possible at present to determine if this movement pattern applies to the whole marginal area as no velocity measurements have been made further upstream. From depth soundings in lake 187 and ice thickness measurements with radar, it is reasonable to believe that the glacier tongue is floating. Water level recordings in lake 187 show that the mean water level is 2 m higher in summer (GTO, 1983). The possibility that the movement pattern is connected with a floating ice tongue and thus is of local origin cannot be excluded.

Table 2. Ice velocity on the margin of the Inland Ice at Påkitsoq

metres per day

Stake	31st May 86 15th Sep 86	15th Sep 86 15th May 87	31st May 86 15th May 87
2.0	0.10	0.05	0.06
2.1	0.15	0.08	0.09
2.2	0.09	0.04	0.06
2.3	0.13	0.08	0.09
2.4	0.10	0.06	0.07
2.5	0.18	0.09	0.12
3	0.30	0.19	0.23
4	0.14	–	–

- Stake not seen. See fig. 2 for stake locations.

Acknowledgements. Bo Madsen, (GTO) helped to survey the stakes, and Peer Jørgensen (GTO) helped with last thermistor readings.

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Mapping and modelling of glacier drainage in the Pâkitsoq basin, central West Greenland

Henrik Højmark Thomsen

Mapping of surface hydrology and modelling of glacier hydraulics at the margin of the Inland Ice north-east of Jakobshavn have been used for investigating glacier drainage. The work is part of the hydropower investigations at Pâkitsoq in a drainage basin proposed for a local hydropower project. Excluding its Inland Ice sector the basin covers an area of 33.6 km^2 and is situated at about 200–600 m a.s.l. (fig. 1; Thomsen, 1988, fig. 1). The main part of the runoff from the basin is meltwater from the ice sheet draining through three lakes, 326, 233 and 187. Lake 187 and 233 are proposed as two separate reservoirs, with tunnels leading to the fjord north-west of the basin.

Glacier hydrological conditions

Meltwater drainage on the ice itself is complicated. Over large areas meltwater drains through innumerable rivers whose drainage courses are influenced by the surface undulation and different structural features on the ice surface. In most cases the rivers escape down into moulins or crevasses, after which the meltwater drainage is controlled by englacial and subglacial drainage conditions. Delineation of drainage basins requires information about supraglacial and subglacial conditions.

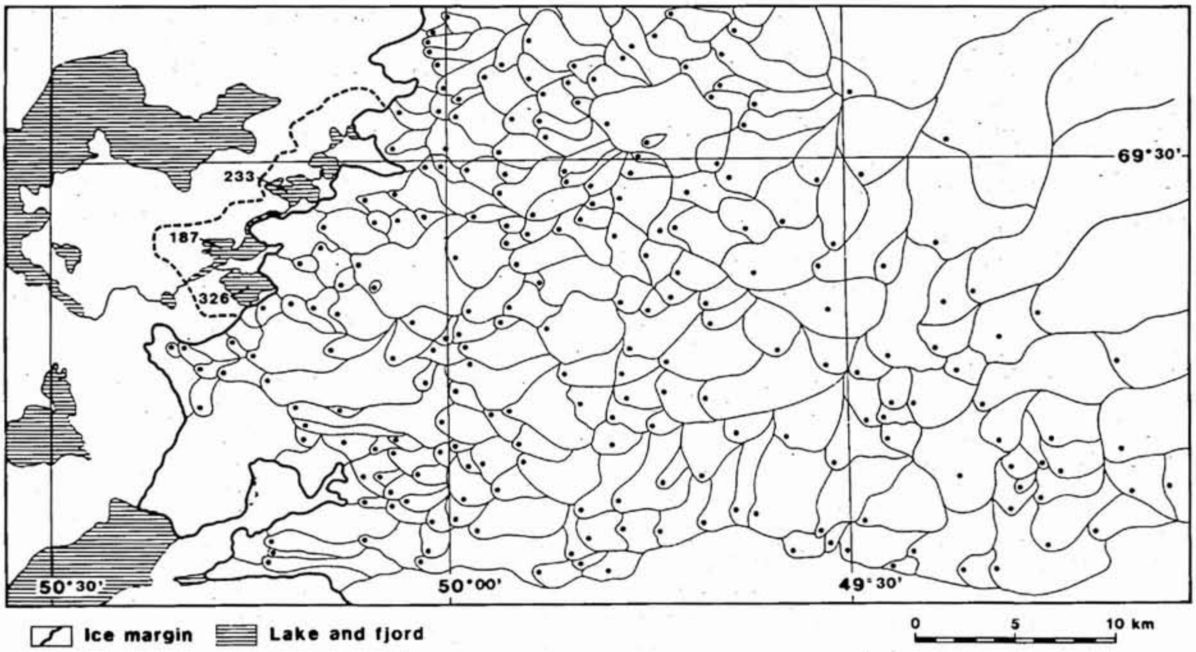

Fig. 1. Drainage cells on the Inland Ice at Pâkitsoq, each draining to a moulin or moulin complex.

Surface topography and drainage

A photogrammetric map on a scale of 1:75 000 was prepared covering the ice-free part of the basin and the adjoining sector of the Inland Ice. The map, based on vertical aerial photographs on the scale 1:150 000 from 10 July 1985, was plotted on a Kern PG-2 stereo plotting instrument connected to a computer system. The map gives physiographic information and surface topography with contour intervals of 50 m in the ice-free area and 20 m on the ice. All possible details have been plotted for the glacier area and trimline zone. This includes features on the ice especially related to surface hydrology such as rivers, lakes and moulins as well as crevasses and lineaments influencing the drainage pattern. The map is supported by observations in the field on foot or from a helicopter.

On the basis of the glacier hydrological map, the ice surface was divided into a number of drainage cells, each draining to a moulin or moulin complex (fig. 1). At the highest elevation, the individual drainage cells have been arbitrarily cut at an elevation of 1100 m a.s.l. The meltwater drainage course from each cell will depend on englacial and subglacial conditions below the point of escape into the ice.

It is reasonable to assume that the drainage pattern on the ice is semi-permanent. It means that drainage takes place in the same valley system on the ice from year to year because the valleys are features of the local topography caused by ice movement over the hilly subglacial terrain. Comparisons with older maps, plotted on vertical aerial photographs from 1959 (Thomsen, 1986) and oblique aerial photographs from 1948 show a stable main drainage pattern with rivers draining in the same valley systems and moulins lying in approximately the same positions. This is despite a mean glacier thinning of 14 m up to an elevation of 500 m a.s.l. in the period 1959–1985.

Englacial and subglacial drainage

Studies of englacial and subglacial water drainage usually deal with drainage in temperate ice conditions. Some water drains through tiny cracks and openings along the single ice crystals (Shreve, 1972), but most water drains in larger channels and conduits starting at moulins and crevasses on the surface. Moulins can reach several hundreds of metres down into the ice, directly to the bottom (Iken, 1972; Meier, 1973). At the bottom of the ice water can drain in a thin water film or in subglacial channels, depending on the regional and local subglacial conditions (Röthlisberger & Lang, 1987).

No direct observations exist about englacial and subglacial drainage from the area at Påkitsoq. The few ice temperature measurements (see Thomsen, 1988) show slightly negative temperatures in the whole ice body. Observations from White Glacier on Axel Heiberg Island suggest that water escaping down into moulins is

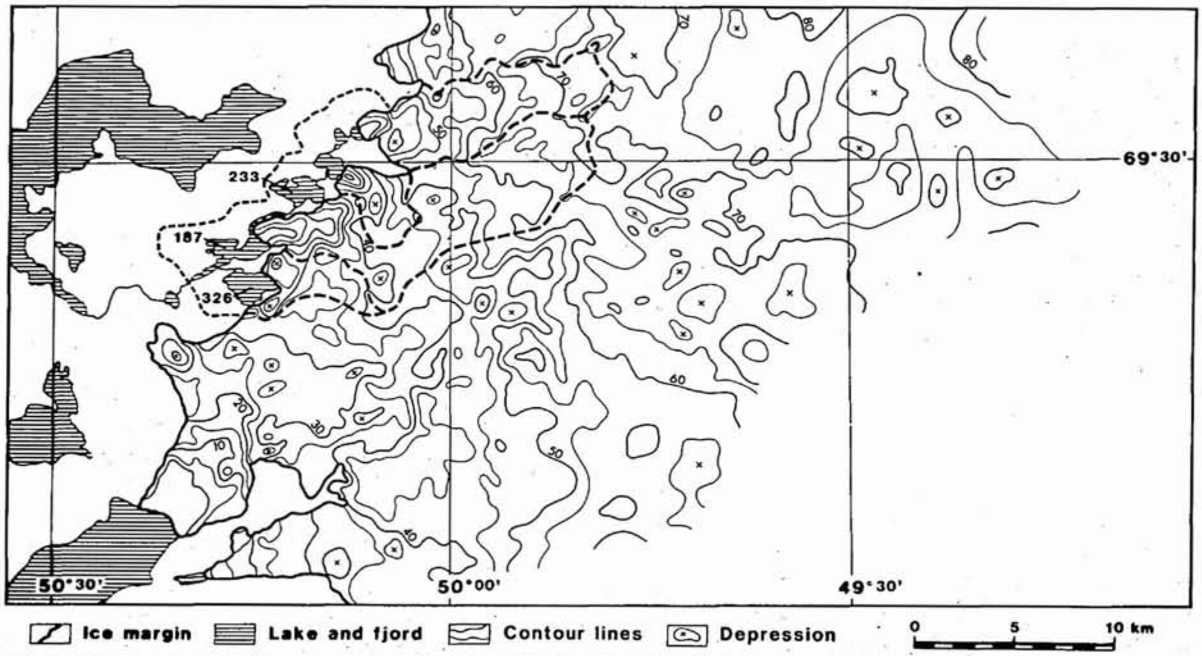

Fig. 2. Calculated subglacial water potential for $k = 0.7$. Units in $10^5 N/m^2$. Subglacial water divides are given by dotted lines.

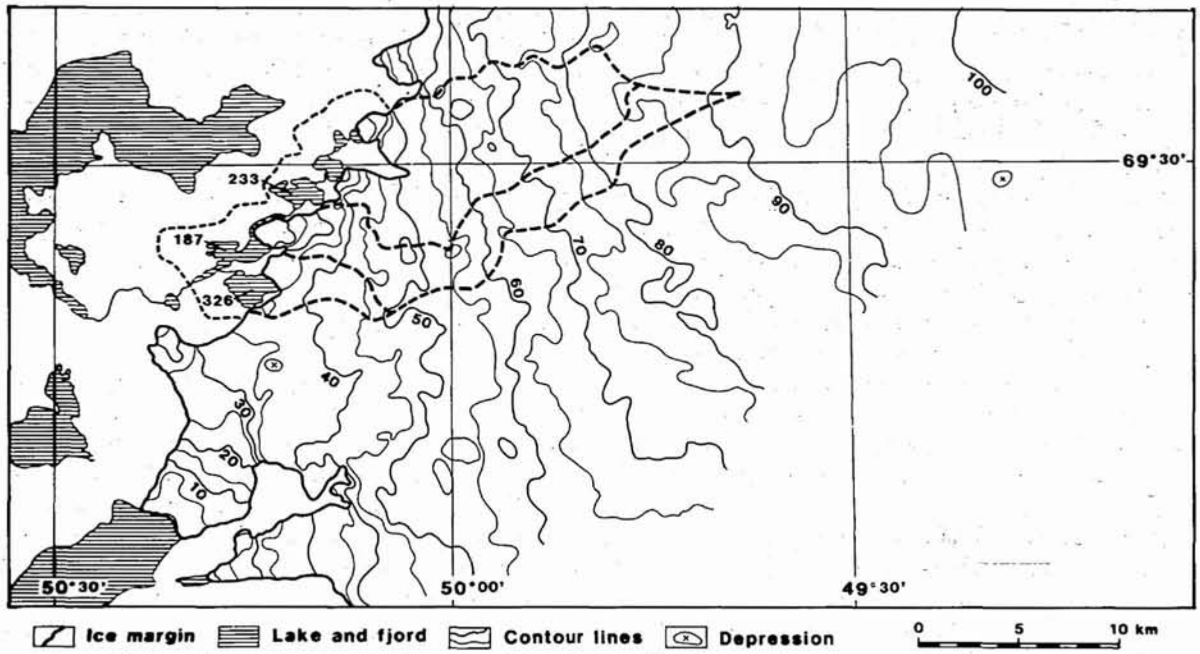

Fig. 3. Calculated subglacial water potential for $k = 1.0$. Units in $10^5 N/m^2$. Subglacial water divides are given by dotted lines

able to reach the bottom of the ice, even under conditions of cold ice (Iken, 1972). From the above it is assumed that surface water at Påkitsoq which escapes down into the ice through moulins quickly flows to the bottom of the ice and there drains according to the subglacial conditions.

Modelling of subglacial drainage

A model by Björnsson (1982) describing subglacial drainage was used for calculating the drainage at the glacier bed. The model describes water drainage down the gradient at a potential which depends upon both surface and subglacial topopgraphy. Data from the photogrammetric surface mapping and data from radio echo sounding of the bed (Thorning *et al.*, 1986; Thorning & Hansen, 1987) were used as input for the model. The potential is expressed as

$$POT = (r_w - k \cdot r_i) \cdot G \cdot Z_b + k \cdot r_i \cdot G \cdot Z_s$$

r_w and r_i are the density of water and ice, respectively. G is the acceleration due to gravity, Z_b and Z_s are the elevation of glacier bed and surface relative to a horizontal datum, and k is a hydraulic factor expressing the relation between subglacial water pressure and ice overburden pressure. Water flowing in an isotropic basal layer drains perpendicular to the potential lines. The model is a first-order approximation of subglacial drainage and does not describe details in water flow.

In practice the k factor will vary in time and with location in a way which is difficult to describe and will, as a minimum, require extensive measurements of the basal water pressure. Subglacial potentials are therefore calculated for a number of k values from zero to one and are plotted as potential maps, some of which are shown in figs 2 and 3.

The subglacial water divides under conditions of varying k factor were drawn from the potential maps. This includes delineation of sub-basins draining to the three separate lakes, 326, 233 and 187. The results show that subglacial drainage is slightly sensitive to changes in the k factor. The drainage area and the sub-basins are nearly constant for $k = 0$ to $k = 0.5$. For $k = 0.7$ there is a slight increase of drainage area in the north-eastern part of the basin and for $k = 1.0$ this tendency is further reinforced, but with a marked change in configuration of sub-basins (figs 2 and 3). The modelling suggests that changes in the subglacial conditions will not create critical changes in the basin size.

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Electromagnetic reflection survey 1987 in key areas of the Pâkitsoq basin at the margin of the Inland Ice, central West Greenland

Leif Thorning and Egon Hansen

The EMR surveys of previous years (Thorning *et al.*, 1986; Thorning & Hansen, 1987) provided the ice-thickness data necessary for a detailed evaluation of the glaciological-hydrological conditions of the Inland Ice required for planning a hydropower plant at Pâkitsoq (Thomsen *et al.*, 1986). This evaluation also identified some locations where ambiguity in the hydrological interpretation resulted in uncertainties in the estimate of water supply to the Pâkitsoq basin, because the data available did not allow the model to predict the direction of drainage of meltwater. Therefore, further geophysical work was undertaken to provide more detailed information on the subglacial relief of these localities on the ice.

Field work 1987

In May 1987, the helicopter-borne electromagnetic reflection (EMR) survey was continued in the Pâkitsoq area. As in previous years a Bell 206 Jetranger helicopter (Glace, OY-HBF) was used. The antenna (a better version in more durable materials) was mounted between the floats, and a new video recorder (Sony VO-6800PS, U-matic format) was used for the recording of EMR data, navigational check marks and navigator comments. A Del Norte line-of-sight navigation system with four remote stations was used for accurate positioning. The remote stations were placed so as to provide optimal coverage in the area of interest and therefore not at the same positions as in 1986. Flight elevation was 10 m above the ice surface. Operations were carried out from Jakobshavn airport with two or three refuelling stops at the camp at Lake 187, Pâkitsoq, where fuel had been landed by a larger helicopter.

The objective of the measurements was to obtain data from two specific areas where problems in hydrological interpretation existed. Thus, from the outset it was intended to utilize the navigation system for feedback of inflight information to navigator and pilot to make it possible to place lines closely and accurately over the critical areas. For a number of reasons this failed in the field, and instead navigation was visual, based on the combined experience of pilot and navigator, and aimed at a concentration of measurements in the areas of special interest. The navigational data were thus only recorded for later processing and were not used in flight.

Processing of data

The navigational data were processed by Bo Madsen, GTO, and transferred to GGU as calculated positions in UTM coordinates correlated with time. Due to errors in the recording of the data in the field, the positioning is less accurate than in 1986. This prolonged the subsequent processing of EMR data because a number of extra checks and corrections have been necessary.

The EMR data were processed using the same methods and computer programs as in 1986 (Thorning & Hansen, 1987). By October 1987 all profiles were reproduced in hard copy, digitized and processed to migrated ice-thickness data. However, the data have not yet been gridded into maps, as it is the intention to compile a combined data base of EMR data from 1985–1987, and scrutinize this for internal consistency before producing improved maps of ice-thickness, subglacial relief and hydrological potential.

Acknowledgements. We thank Bo Madsen, GTO (cooperation in the field); Ove Karlquist, Greenlandair Charter pilot (flying under difficult conditions); M. Svane Jørgensen and Dorthe Nyland Sørensen, GGU, (data processing); Anders Clausen (construction of the new radar antenna). Part of the cost of GTO's navigation system was defrayed by Energiafdelingen, GTO.

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Construction and testing of a lightweight radar for ice-thickness determinations on glaciers in the Pâkitsoq area, central West Greenland

Leif Thorning and Egon Hansen

Although the helicopter-borne EMR surveys described in Thorning & Hansen (1988) resulted in excellent results from most of the catchment area of interest for the hydropower plant planned for Jakobshavn, there are still some local areas, such as the glacier lobes and ice-falls, where no meaningful data could be obtained. In the autumn of 1986 it was decided to construct a monopulse ice radar for use in such areas.

Construction

The instrument was constructed at GGU using a subcontractor for the high voltage power supply. The principle of the ice radar and good advice on its construction is given in Hodge (1978), which we followed closely. The principle and main components are shown in fig. 1.

The instrument consists of two separate parts: the transmitter and the receiver.

The transmitter contains a 750 V DC power supply and a circuit containing four transistors. These are brought to avalanche and pulse with a repetition rate of approximately 6 kHz over a load of 400 ohm. The choice of output frequency is then made through the choice of antenna length.

The signal from the receiver antenna is fed into an oscilloscope. The signal on the oscilloscope shows the direct pulse through air (used for triggering the sweep) and the reflected pulse delayed in proportion to the distance travelled through ice (fig. 2). The two-way travel time is calculated as the delay time for the arrival of the reflected pulse read off the oscilloscope plus the travel time for the direct pulse through air. For documentation a photograph is taken of the oscilloscope screen and its setting using a camera with date/time backpanel.

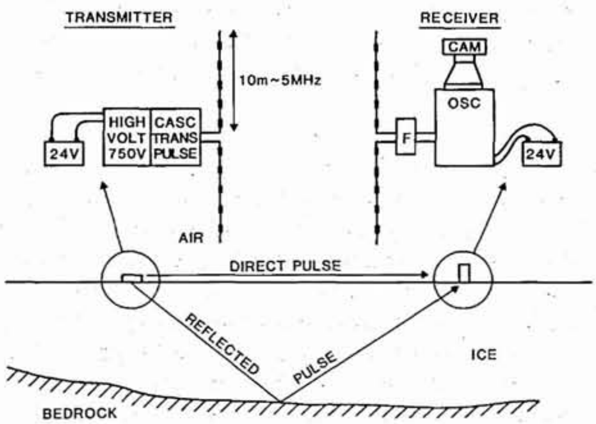

Fig. 1. Principle of the mono-pulse ice radar. For explanation, see text.

Several sets of antennas were constructed for different frequencies, in half-antenna lengths of 1.25, 2.5, 5, 10, 20, and 40 m, corresponding to peak frequencies of 40, 20, 10.5, 2.5, and 1.25 mHz. Each antenna is a resistively-loaded, centre-fed dipole, identical and symmetric about the feed point. The antenna wire with interspersed resistors was enclosed in a plastic tube with test points at the ends to check for continuity. The antennas were flexible and easy to handle even at the low temperatures prevailing on the ice. The power supply was from rechargeable batteries for both transmitter and receiver.

Velocities for radio waves were chosen at 300 m/μs in air and 168 m/μs in ice.

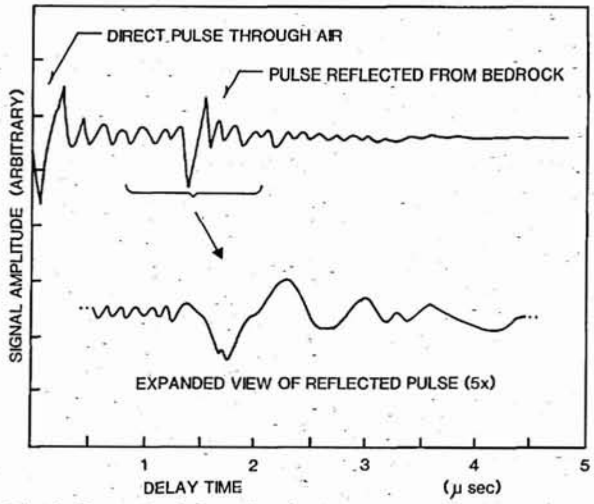

Fig. 2. Example of recording (redrawn from photo).

Field work 1987

As part of the glaciological work in the Pâkitsoq area (Thomsen, 1988) and the continuation of the EMR surveys, the last of which took place in April 1987 (Thorning *et al.*, 1986; Thorning & Hansen, 1987; Thorning & Hansen 1988), a program was planned for detailed measurements around drill sites and on the glacier lobes to test the instrument and produce useful data for the hydrological work. Besides using the ice radar at different localities and on different types of ice, a series of experiments using different frequencies, different transmitter-receiver distances, profiling and expanding spread around a fixed common midpoint were planned to demonstrate the capabilities and limitations of the equipment on the Greenland Inland Ice. However, as it turned out, much time was wasted on equipment failures manifesting themselves in burned-out components in the high voltage supply circuit. Thus, only a limited part of the program could be carried out. It was, however, enough to demonstrate the usefulness of the ice radar, and to get some ideas on how to improve GGU's version. The work was based at the camp on the Inland Ice shared with the GGU glaciologist working in the area (Thomsen, 1988).

Discussion of results

Work was carried out in three areas.

In the area adjacent to the first hole drilled through the ice (Olesen & Clausen, 1988), the ice thickness of approximately 300 m fitted well with both the results from the EMR surveys and from the logging of the hole. The returned signal was weak, but easily recognizable.

The second area was near a drainage system in a depression in the ice surface where the drilling ran into difficulties, probably due to material of some sort in the ice at approximately 270 m depth and some 60 m above the bedrock. Here transmitter-receiver (T-R) distance was varied from 50 to 200 m and all frequencies were tried. Although the direct pulse was always well defined, no reflected pulse could be recognized in the field. Presumably, the material (sand, gravel, rock ?) which gave difficulties for the drilling, also completely scattered the radar waves and diminished the returning signal to below the noise level. This does not mean that it is impossible to obtain data from such an area in the future. The 'contamination' of the ice may vary locally,

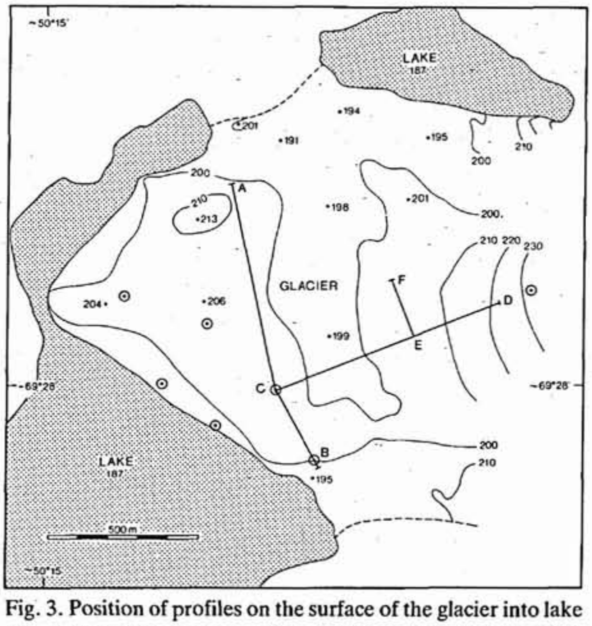

Fig. 3. Position of profiles on the surface of the glacier into lake 187. Transmitter-receiver distance 100 m. The circled points are positions of stakes. Contours and points show elevation of surface.

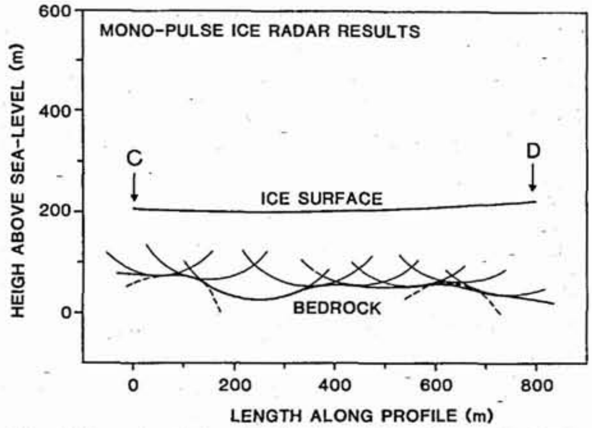

Fig. 4. Example of plot of reflection ellipses for profile CD in fig. 3. The two tops (at 100 and 600 m) would be an equally correct interpretation (shown with stippled line).

and it may be a question of moving the equipment around, until a place is found where the signal gets through. Unfortunately, there was no time to do this in 1987.

The third area, the lobe of the glacier into lake 187, yielded the most promising results. Here 2 km of profiles with a T-R separation of 100 m gave good, clear return signals at all measuring points. The antenna length was 20 m corresponding to 5 mHz, a frequency often quoted in the literature as suitable for this type of environment. In fig. 3 the position of the profiles on the glacier is shown where point C is at stake 2.1.

The calculated two-way travel time cannot be directly converted into ice thickness, because it cannot be assumed that the return signal comes from below the midpoint. For each position of the transmitter and receiver the reflected signal may come from any point on an ellipsoid with transmitter and receiver in the focal points. A program has been made following Blatter (1987; personnal communication, 1987), which calculates and plots this ellipse in the vertical plane of the profile for each measurement position (fig. 4). If no other information is available, the envelope of these ellipses probably represents the best estimate of the upper limit of the bedrock relief. However, alternative interpretations are possible, as indicated on fig. 4. More detailed measurements using other geometries and sampling intervals may well limit the ambiguity significantly.

Using this simple approach the cross-sections (fig. 5) were constructed giving ice thicknesses of 70–50 m. These represent the first estimates of the ice thickness of this glacier based on direct measurements, but they are not yet sufficiently detailed to determine the character of the bottom of the glacier, i.e. whether it is a

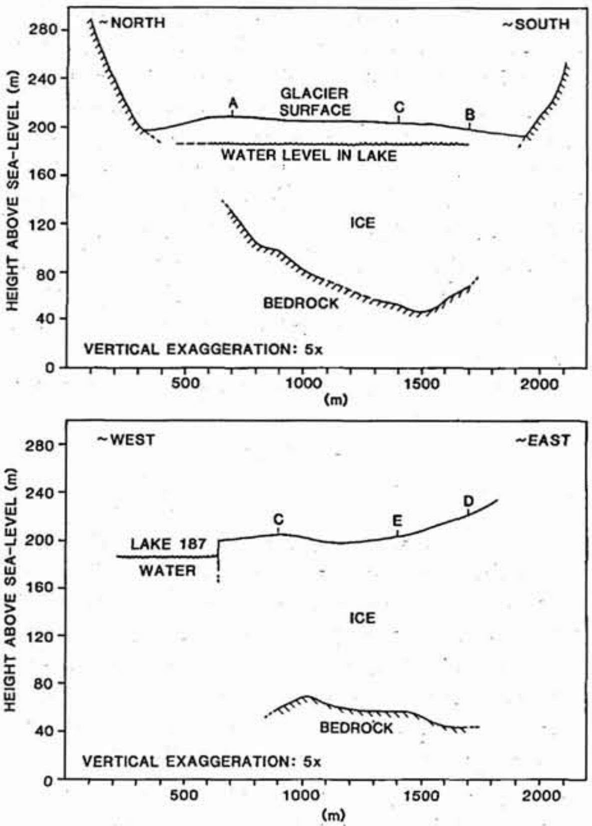

Fig. 5. Sections constructed from the ice radar measurements. Profile positions indicated on fig. 3.

bedrock/ice or water/ice interface; at present the former possibility is favoured. The results will be added to the EMR data base, and improve map accuracy in this area.

Future work

Based on the experiences with the monopulse ice radar, technical improvements have been planned. More detailed work will probably be attempted next year on the lobe of the glacier into lake 187, aimed at an accurate mapping of the bedrock/water surface below the glacier. Methods for data reduction will also be refined.

Acknowledgements. We would like to thank other members of the camp on the ice for pleasant cooperation. Lisbeth Christensen, GGU, participated in this work as an assistant.

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Test drilling with a hot water jet at the Inland Ice margin, Pâkitsup, central West Greenland

Ole B. Olesen and Anders Clausen

A new hot water jet drill was tested on the Inland Ice margin north-east of Jakobshavn. A total of 2436 m was drilled and data on drilling performance is presented together with the preliminary interpretation of borehole logging.

Introduction

One of the main problems in assessing potential water sources for hydroelectric power plants using meltwater from the Inland Ice is the delineation of the drainage basin. In the case of the proposed project at Paakitsup Akuliarusersua this delineation has been attempted by a combination of surface topography and drainage systems, radio-echo soundings, mass balance measurements, ice dynamics and present theories on the internal drainage of glaciers (summarized by Thomsen *et al.*, 1986).

The subglacial or englacial drainage of glaciers, particularly where temperatures are below the pressure melting point is not very well understood. From both practical and scientific considerations it therefore seemed reasonable for GGU to start research using the basin at Paakitsup Akuliarusersua as a test area. Among important aspects of this research are the monitoring of hydrostatic pressure variations within the drainage system, tracer experiments and subglacial and surface topographic mapping. This requires the installation of gauges at different levels in the ice, the injection of tracers and withdrawal of water from specified locations and spot checks on the accuracy of the radio-echo soundings made in the area. The only practical way of accomplishing this programme is by drilling, and it was therefore decided that GGU should either buy or develop its own drill.

Equipment

As no ice-core samples were required and a regular borehole diameter was unimportant, it was decided to use a hot water jet to penetrate the ice. This type of drilling is the fastest and most reliable method and has been widely used in the Alps (Iken *et al.*, 1977). We had an opportunity to examine the Swiss system at the Swiss Federal Institute of Technology (ETH) in Zürich and used it as a basis for our own drill.

The basic principle consists of pumping water through a heating system and into a hose with a rigid drill tip with a nozzle. The hot water from the nozzle melts the ice in front of it and flows back up the drill hole. All the main elements in the system are commercially available, but the actual fitting and mounting of the different components is unique to this system. The central part of the drill consist of two units, a power unit and a heating unit (fig. 1). The power unit is a 6 kW four stroke gasoline engine with reduction gear connected with an 0.4 kW 220 V generator and a piston pump capable of delivering 18 l/min. of water with a pressure of up to 100 bar. Fuel consumption is 2.5 l/hr and the total weight of this unit is 79 kg. The heating unit is a water circulation coil with an oil burner (modified for using jet A1 fuel). The airblower and oil pump are driven by a flexible axle from the power unit, and current for the ignition is drawn from the generator. Fuel consumption is 11.6 l/hr corresponding to 113 kW and the weight is 127 kg. During drilling the two units can be used separately (to

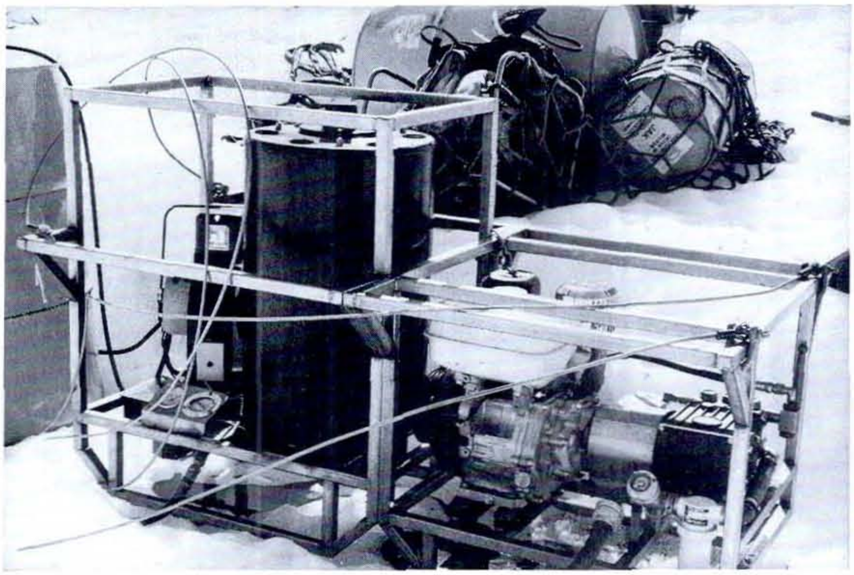

Fig. 1. Components of the drilling system. To the right, a gasoline engine and high pressure pump form the power unit. The heating unit is on the left. Units are bolted together for field operation.

the extent of the flexible axle) but will normally be bolted together to work as a single unit.

Further equipment used is 500 m of heat resistant high pressure hose (working temperature and pressure are 121°C and 138 bar) in lengths of 100 m, a 2 m long 25 mm diameter stainless steel drill tip with interchangeable nozzles and a clinometer on top, a lightweight tripod with winch and pulley, and a low pressure pumping unit for use when water has to be drawn from farther away.

The total weight of the complete drilling system inclusive of tools and spare parts is 435 kg.

Field tests

Before the actual deep drillings a few tests were made to determine the size of nozzle to be used. It was soon discovered that with the given weight of just under 5 kg for the drill tip, nozzles with an inner diameter of 2.4 mm or less forced the tube back up the drill hole. With a 2.5 mm nozzle it was impossible to steer the drill vertically, so in the end a 2.7 mm nozzle was chosen. This gives a pump pressure of 35 bar with one 100 m length of hose and all subsequent drillings were made with this nozzle.

Eight deep holes were made to depths between 270 and 383 m totalling 2,436 m (see Table 1). In two areas, two holes were drilled 10 to 15 m apart, while in the other areas only one hole was made (for site location, see map in Thomsen, 1988). Of the eight holes, five went to the bottom while two (nos 1A and 6) are uncertain and one definitely stopped in the ice (no. 3A). The outlet temperature from the heating unit was just about 82°C, giving a heating efficiency of above 90%. To get an idea of the heat loss in the hose a thermistor was mounted in the drill tip and exposed to the circulating bore-hole water.

The measured temperatures are plotted in fig. 2 against length of hose between heater and thermistor housing. Typically about 20 m of hose was above the water-filled drill hole when temperatures were measured. As only a few data points are available a straight line has been fitted to the points. The data are, however, consistent with a heat loss of 2% per 10 m of hose which is very close to the 1.8% measured in an experiment in Copenhagen prior to the field tests.

In Table 1 measured drilling speeds in m/min. are

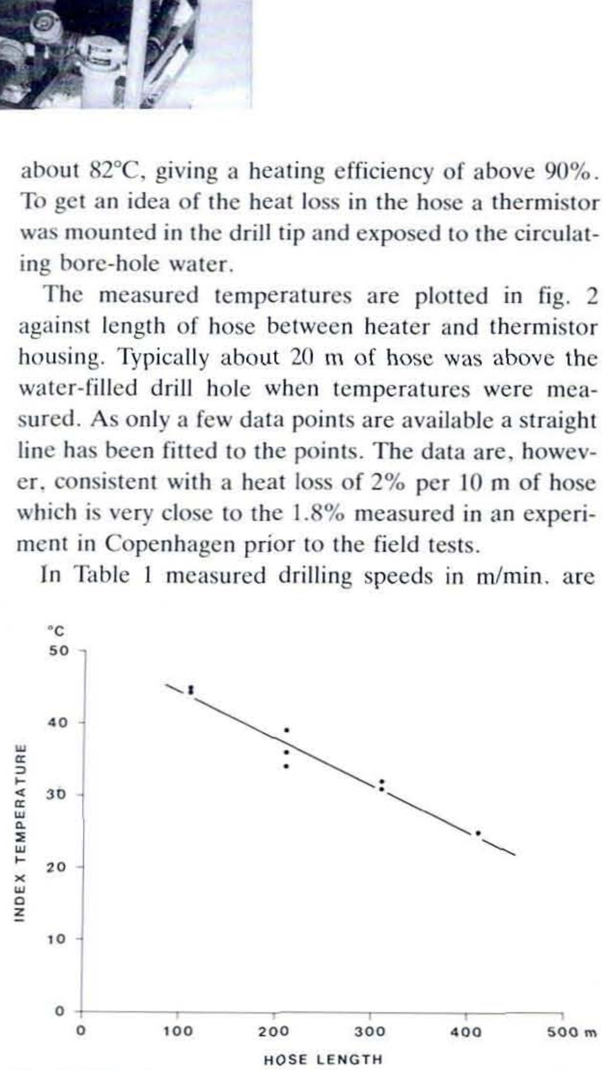

Fig. 2. Fall in temperature with increasing hose length. Temperatures are read with about 20 m of hose above the hole. For explanation of index temperature and thermistor setting see text.

Table 1. Drilling speed and depth for hot water jet Paakitsup Akuliarusersua, Jakobshavn, West Greenland August 1987

Drilling speeds in metres per minute.

Site	Depth intervals in metres				Max. depth m
	0–90	90–190	190–290	>290	
8	4.1	2.6	2.2	–	298
3	6.0	–	–	–	305
3A	–	–	–	–	275*
2	5.6	–	–	–	270
2A	–	–	–	–	278
1	6.4	3.6	2.1	1.1	382
1A	4.5	4.0	–	–	298*
6	–	–	–	–	330*
Mean drilling speed	5.2	3.3	2.1	1.1	2436 Total

* Drilling did not reach bottom of glacier

listed for various intervals and the same figures are plotted in fig. 3. The actual measurements refer to times between fitting successive 100 m lengths of hose. No corrections have been attempted to compensate for 'difficult' drilling conditions when the drill penetrated layers of ice with debris which had a considerable slowing effect.

All the listed drillings in Table 1 were carried out with the same nozzle which means that the pressure at the tip of the drill was constant. However, on one occasion a large nozzle (3.0 mm) was tried at a depth of 200 m and the drilling speed dropped to about one third of that with the 2.7 mm nozzle. Although this was the only trial it indicates that pressure has a very strong influence on drilling performance.

Comments

In general, drilling proceeded at an even speed but at sites 3 and 3A a very resistant layer was encountered 220 m below the surface where drilling speed dropped to almost nil for 10 to 20 m. Subsequent inspection of the nozzle showed distinct scratches along the sides but no fine particles were observed in the upwelling water in the drill hole. Most probably the layer encountered was a shear plane carrying material from the bottom upstream of the drill site. Downstream from the site shear planes could be seen on the surface although these seemed to carry a relatively high content of fine particles of silt and clay sizes.

At site 1A a deep layer carrying debris was encountered and a sample was unwittingly brought to the surface. In order to read the clinometer, water pressure was taken off the drill by suddenly turning the relief valve. The subsequent sudden pressure drop must have created a vacuum in the hose (probably due to the elastic stretching and contraction) and the nozzle became clogged with sand. The individual grains are up to 2.5 mm in diameter and are mostly sharp edged and hence have not been transported for long distances. The samples are being mineralogically examined for composition and possible provenance. The hole was not finished to the bottom because of refreezing in the hole while the nozzle was being changed and examined for damage.

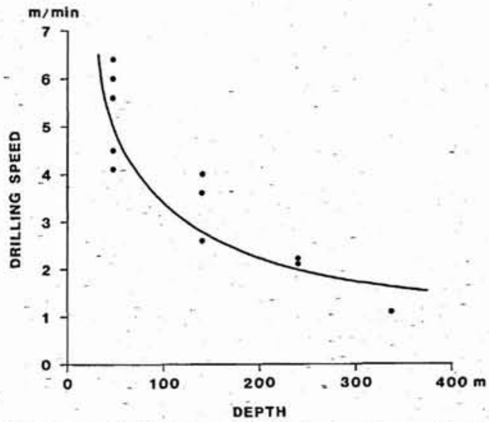

Fig. 3. Decrease of drilling speed with increasing depth. Same data as in Table 1.

At drill site 1 the drill seemed to hit a void at 180 m below the surface when the water level in the drill hole suddenly dropped. When the drilling stopped, the water level rose again and stood at 20 m below surface, that is when an additional 2.5 m^3 of water had been pumped into the hole.

In Table 2 ice thickness measured by both radio-echo soundings (EMR) and by drilling with the hot water jet is compared. EMR thickness is read by interpolation between computer-generated contours based on a 100

Table 2. Comparison of ice thickness measured by radio-echo soundings (EMR) and hot-water jet drill holes

Site	Drilling m	EMR m	Difference m	%
8	298	300	2	0.7
3	305	300	5	1.7
2	270	240	30	11.1
2 A	278	240	38	13.7
1	382	400	18	4.7

m grid of filtered data. Sites 2 and 2A are only 10 m apart and the difference of 11% and 14% in the two data sets are probably due to the great bottom gradient of approximately 40° which both aggravates positioning errors and influences EMR interpretation.

Although the overall performance of the drilling system was very satisfactory, a few things must be changed to improve determination of when the bottom has been reached. For instance, the present clinometer is too sensitive to vibrations so that the pump has to be stopped during the 'feeling' for the bottom. A heavier drill tip should be used both for the advantage of drilling with a smaller diameter nozzle and for the bigger loss of weight when the tip rests on the bottom, making it easier for the drilling crew to detect when to stop paying out more hose.

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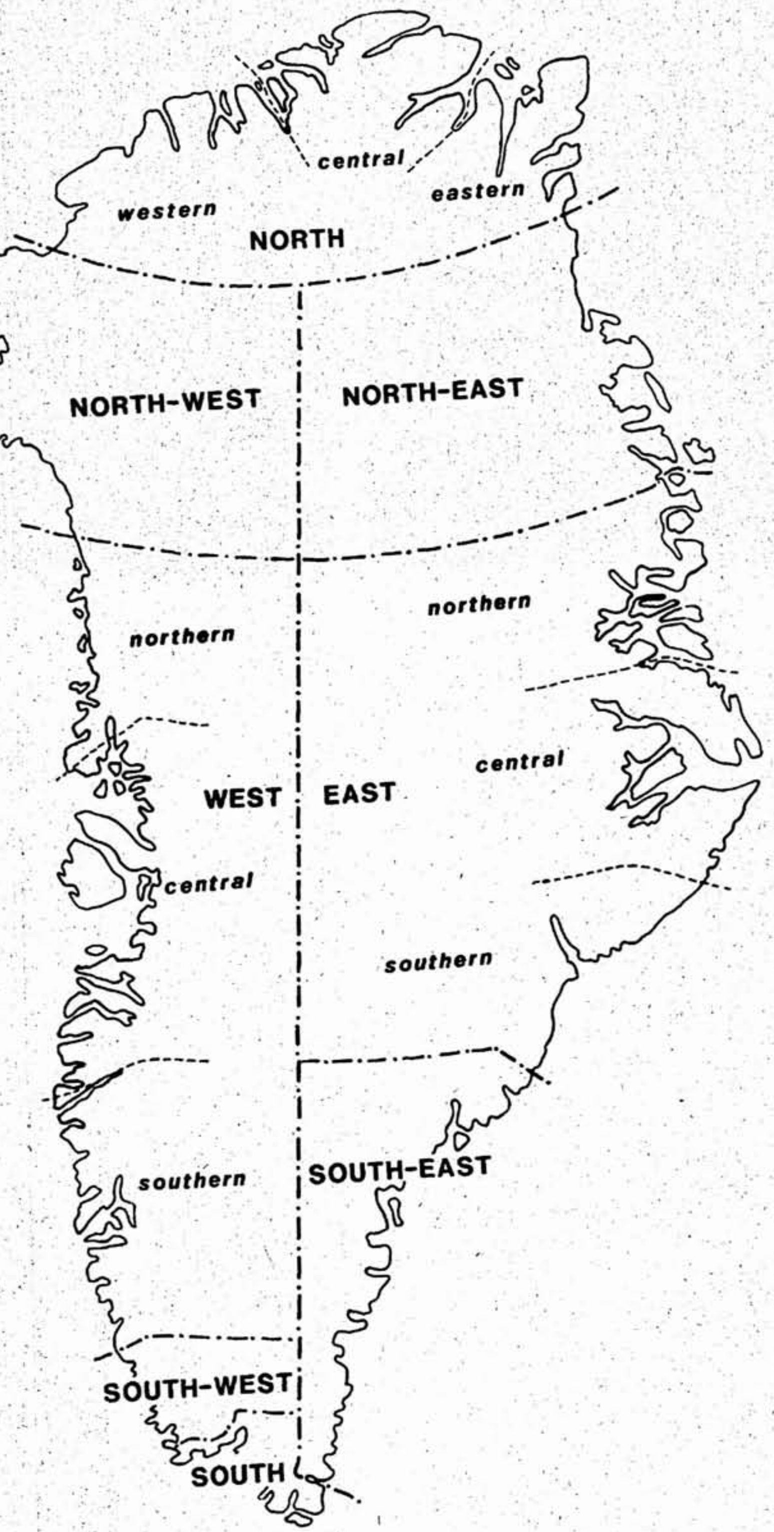

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