Magmatic accretion of grey gneiss precursors, contemporaneous structural evolution, and late-tectonic granite domes and sheets

Outline of the main orthogneiss units and their structural and metamorphic evolution

Prior to systematic work in the Fiskefjord area in the 1980s the orthogneisses of the Akia terrane north of Godthåbsfjord were only poorly known; an overview of previous work was given in the introduction. Allaart (1982) made a general distinction between amphibolite facies and granulite facies (enderbitic) rocks, and recognised two large homogeneous tonalite complexes north and north-east of Fiskefjord, the Taserssuaq tonalite and Finnefjeld gneiss complexes. Earlier, Macdonald (1974) and especially Reed (1980) had studied dioritic and leucocratic gneisses in Nordlandet. During the subsequent systematic survey of the Fiskefjord area in the 1980s the quartzo-feldspathic rocks of intrusive magmatic origin were resolved into several groups of progressively younger and more leucocratic rocks which are described in the following sections.

The *c.* 3220 Ma Nordlandet dioritic gneiss (ion probe zircon age data shown in Fig. 25 and discussed below), which contains enclaves of an even older supracrustal association, is the earliest nucleus of continental crust that has been identified in the Akia terrane (Table 1). It forms most of Nordlandet south of the Fiskefjord area, and underlies large tracts in the southern part of the Fiskefjord area (Plate 1). The remainder of the grey gneiss is dominated by *c.* 3000 Ma old grey, K-poor, amphibolite facies tonalite-trondhjemite-granodiorite (TTG) orthogneisses and their brown granulite facies equivalents, and also comprises younger dioritic rocks. Besides there are local sheets of mesoperthite-bearing (granulite facies) granitic gneiss. The orthogneiss protoliths were emplaced into the older assemblage of supracrustal rocks and Nordlandet dioritic gneiss, and deformed and metamorphosed during a major event of continental crustal accretion at *c.* 3000 Ma.

Most orthogneiss precursors intruded as sheets of variable dimensions. Their emplacement was accompanied by regional deformation, which comprised at least two phases of isoclinal folding followed by a phase of upright folding and development of N–S trending straight zones contemporaneously with granulite facies metamorphism (see p. 41). Two large dome-like massifs were emplaced in the area. The Taserssuaq tonalite

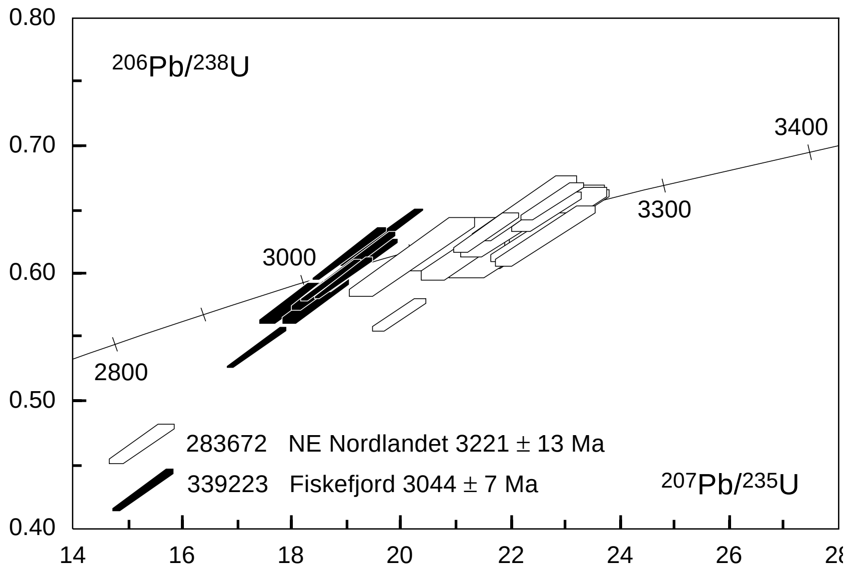

Fig. 25. SHRIMP U-Pb zircon data from Nordlandet dioritic gneiss, GGU 283672 (3221 ± 13 Ma, with a number of younger ages at *c.* 3180 Ma) and Qeqertaussaq diorite, GGU 339223 (3044 ± 7 Ma). Sample localities are shown on Fig. 52 (analyst A. P. Nutman, using Sri Lankan zircon SL13 as standard. See Compston *et al.*, 1984 and Friend & Nutman, 1994 for details of the analytical procedure). The zircons in sample 283672 have very low U contents of 15–50 ppm, but the analysed crystals were large enough that a normal count rate could be obtained with a large beam diameter (*c.* 25μ). The 3221 ± 13 Ma age is derived from a weighted mean of ^{207}Pb/^{206}Pb ratios, using analyses representing the least isotopically disturbed sites in the same morphological type of zircons.

complex was emplaced syn-granulite facies metamorphism, as its western parts display evidence of retrogression from granulite facies parageneses; most of the complex occurs north-east of Fiskefjord where granulite facies conditions were never reached at the exposed level. Second, the Finnefjeld gneiss complex to the north-west appears to entirely post-date granulite facies metamorphism in that area. The Igánánguit granodiorite and the Qugssuk granite (Garde *et al.*, 1986), two suites of granitic rocks in the north-eastern and eastern part of the Fiskefjord area, were formed by melts generated from older orthogneiss and emplaced after the granulite facies and main deformation events.

A brief overview of the Archaean metamorphic evolution of the Fiskefjord area, described in detail later, follows here as introduction to the field descriptions. Granulite facies metamorphism occurred at *c.* 3000 Ma over most of the Fiskefjord area, except its eastern and northernmost parts. The granulite facies metamorphism was thermal and probably caused by heat accumulated during the continuous injection of tonalitic magma into the growing continental crust (McGregor *et al.*, 1986; Garde, 1990), a mechanism that had previously been suggested by Wells (1979, 1980) for the neighbouring Buksefjorden area. The central and northern parts of the Fiskefjord area were subsequently partially or completely retrogressed from the granulite facies. Most of the orthogneiss in these areas has disequilibrium amphibolite facies mineral textures, whereas the mafic to ultramafic rocks of the supracrustal association are much less affected by the retrogression. Garde (1989a, 1990, 1991) showed that both granulite facies metamorphism and retrogression were accompanied by migration of particularly LIL elements in the orthogneisses, and suggested that much of the retrogression took place very soon after the culmination of granulite facies metamorphism. The pervasive retrogressive event was followed by younger, at least in part Proterozoic retrogression localised along narrow ductile shear zones, along brittle faults, and along the margins of mafic dykes.

Grey orthogneiss, Taserssuaq tonalite and Finnefjeld gneiss complexes

Dioritic gneiss

Dioritic, quartz-dioritic and mafic tonalitic gneiss (shown as dioritic and mafic dioritic gneiss by Garde, 1987, 1989b) constitutes most of the south-western part of the Fiskefjord area and continues southwards into the Nordlandet area (McGregor, 1993, fig. 5); in the following it is referred to as the Nordlandet dioritic gneiss.

Sensitive high-resolution ion microprobe ('SHRIMP') U-Pb analysis of very clear, pale pinkish, euhedral to slightly corroded zircons in a sample of dioritic gneiss collected in north-eastern Nordlandet (GGU 283672) gave a ^{207}Pb-^{206}Pb age of 3221 ± 13 Ma (Fig. 25); the analyses were performed by A. P. Nutman at the National University of Australia. A cluster of younger ages was also obtained from morphologically similar zircons in GGU 283672, suggesting a thermal event at *c.* 3180 Ma. A whole rock Pb-Pb isochron age of 3112 ± 40 Ma, based on samples of both Nordlandet dioritic gneiss, amphibolite facies grey gneiss and Igánánguit granodiorite, was previously published by Garde (1989a); regression of the dioritic gneiss samples alone gives a similar age with a much larger uncertainty. The *c.* 3220 Ma protolith age of the Nordlandet dioritic gneiss is supported by data from the southernmost part of the Akia terrane; A. P. Nutman and H. Baadsgaard (personal communication, 1995) obtained SHRIMP zircon ages of 3235 ± 9 Ma and 3193 ± 7 Ma from two samples of dioritic gneiss collected on the mainland 5 km north-east of Nuuk, and in southern Nordlandet *c.* 8 km north-west of Nuuk. Zircons in the latter sample contain metamorphic overgrowths dated at 3014 ± 7 Ma.

In the south-western part of the Fiskefjord area the Nordlandet dioritic gneiss appears to form three *c.* 10–15 km wide N–S trending bodies, which are separated by 1–4 km wide linear tracts of tonalitic orthogneiss and amphibolite that probably represent the cores of steep, isoclinal folds (as suggested by fold closures to the south, see McGregor, 1993, fig. 5); the latter rocks may originally have been intercalated with the dioritic gneiss along early thrusts. The northern part of the Nordlandet dioritic gneiss, south of central Fiskefjord, is intruded by, and folded together with tonalitic orthogneiss in a complex fashion (Garde *et al.*, 1987). Smaller bodies of dioritic gneiss, which may or may not be part of the *c.* 3220 Ma Nordlandet dioritic gneiss, occur in most other parts of the Fiskefjord area (an amphibolite body north-east of Sangmissup nunâ (Plate 1) was incorrectly shown as dioritic gneiss by Garde, 1989b). Where primary field relationships are preserved, the dioritic gneiss is intruded and commonly agmatised by more leucocratic gneiss. The diorite itself can locally be observed to truncate supracrustal amphibolite, but enclaves of supracrustal amphibolite in the dioritic gneiss have only locally been observed; they are difficult to distinguish on weathered outcrops of granulite facies rocks and may therefore be much more

Fig. 26. Granular, medium-grained granulite facies metadiorite with indistinct quartz-plagioclase migmatitic network. 2 km west of central Quagssûp taserssua.

Fig. 27. Partially retrogressed granulite facies dioritic gneiss with vertical schistosity and rare leucocratic veins. N–S trending dioritic unit, 2 km due north of the head of Fiskefjord. The penetrative schistosity was developed before or during granulite facies metamorphism, and subsequent growth of retrograde amphibole and biotite was not accompanied by renewed deformation. The locality lies along the northern extension of the Qugssuk–Ulamertoq zone (Fig. 44) and shows that the development of this N–S trending high strain zone preceded retrogression.

common than hitherto recognised. In certain areas trains of rounded, up to metre-sized, fragmented gabbroic and leucogabbroic enclaves (Fig. 14) are readily distinguished in the field by their specks of white calcic plagioclase. Some of these enclaves can be traced for many kilometres and outline large refolded folds (McGregor, 1993; see p. 41).

Fresh outcrops of granulite facies dioritic gneiss may locally display several intrusive phases (Garde, 1990), but it typically appears as a very homogeneous, dark, brownish weathering, medium-grained and granular rock with a weak quartzo-feldspathic network (Fig. 26). South of outer Fiskefjord the dioritic gneiss commonly possesses a weak to distinct schistosity formed by platy quartz and elongate aggregates of mafic minerals with equilibrium granulite facies textures. In this area the schistosity is in places folded on metre-scale, south-plunging, upright to overturned folds probably belonging to the Pâkitsoq phase (see p. 41). Figure 27 shows another example of dioritic gneiss with vertical (granulite facies) schistosity from the northern extension of the N–S trending Qugssuk–Ulamertoq zone, which is described below.

A geochemically distinct group of dioritic gneiss, the Qeqertaussaq diorite (see p. 51), is restricted to the peninsula Qeqertaussaq and its vicinity in central Fiskefjord, where it forms common enclaves of very variable size in tonalitic-trondhjemitic gneiss (Fig. 28; see also Plate 1 and Garde, 1989b). In the field the Qeqertaussaq diorite is indistinguishable from dioritic gneiss

Fig. 28. Enclave of dioritic gneiss (Qeqertaussaq diorite) in tonalitic orthogneiss. The rocks are retrogressed, with development of blebby texture. North-eastern end of Tasiussarssuaq. Photo: S. B. Jensen.

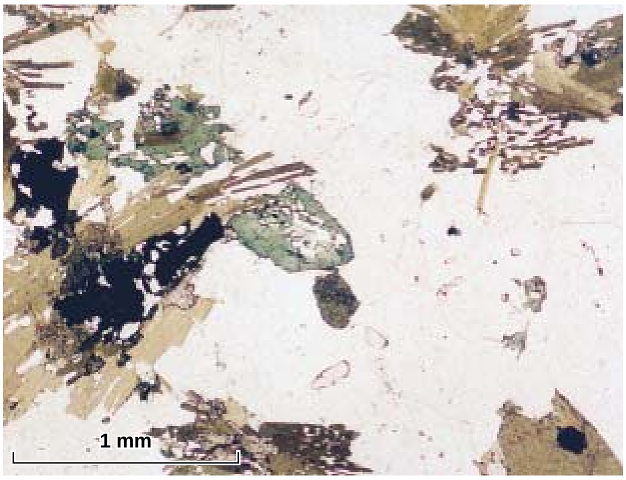

Fig. 29. Thin section of retrogressed dioritic gneiss (Qeqertaussaq diorite) with typical blebby texture: aggregates of spongy blue-green amphibole (after orthopyroxene) and sheaves of olive-brown biotite. GGU 328527, 5 km west-south-west of Kûlik.

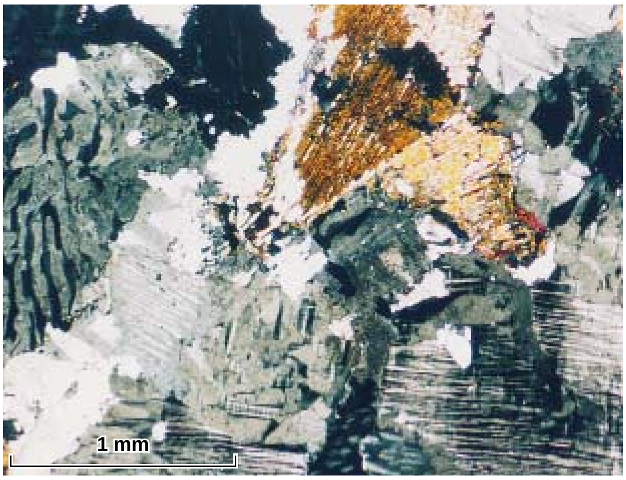

Fig. 30. Thin section of partially retrogressed dioritic gneiss (crossed polarisers), showing mesoperthite feldspar and clinopyroxene with secondary biotite overgrowths. The preservation of mesoperthite (exsolved from granulite facies intermediate alkali feldspar) shows that recrystallisation during the retrogression was incomplete. GGU 328583, east coast of island at the head of Tasiussarssuaq.

in other parts of the Fiskefjord area. The Qeqertaussaq diorite is considerably younger than the Nordlandet dioritic gneiss and is an early member of the main phase of grey gneiss; a SHRIMP Pb-Pb zircon age of 3044 ± 7 Ma was obtained from a sample (GGU 339223) collected opposite Qeqertaussaq at the south coast of Fiskefjord (Fig. 25; analyst A. P. Nutman). The zircons in this sample are light brown, euhedral or with partially rounded terminations, without obvious cores or rims, and contain 100–200 ppm U.

Granulite facies dioritic gneiss mainly consists of plagioclase, quartz, hypersthene, hornblende, iron-titanium oxides and commonly diopsidic clinopyroxene; fox-red biotite is also quite common. Textures in thin section suggest all ferromagnesian minerals to be in metamorphic equilibrium. Part of the dioritic gneiss is partially or completely retrogressed and contains irregular intergrowths of quartz and secondary blue-green amphibole, which commonly has lower contents of alkali elements and titanium than the granulite facies hornblende (Garde, 1990; see p. 59). Sheaves of secondary green or light brown biotite are likewise low in titanium. In thin section the Qeqertaussaq diorite resembles other dioritic gneiss members, except that appreciable amounts of K-feldspar may occur and accessory apatite is much more common. Figures 29 and 30 display typical disequilibrium retrogression textures in Qeqertaussaq diorite.

Fig. 31. Amphibolite facies biotite gneiss with tight folds. In spite of strong deformation, the textural details of polyphase magmatic emplacement have been preserved. Coastal outcrop at the north-eastern end of Bjørneøen.

Fig. 32. Retrogressed tonalitic gneiss with irregular quartzo-feldspathic network. Patches of secondary amphibole and biotite preferentially occur along the leucocratic segregations. Small island in inner Fiskefjord, 3 km south-east of Kûlik.

Fig. 33. Polyphase, partially retrogressed orthogneiss with blebby texture developed both in the leucocratic and more mafic parts of the rock. 7 km north-east of Ulamertoq.

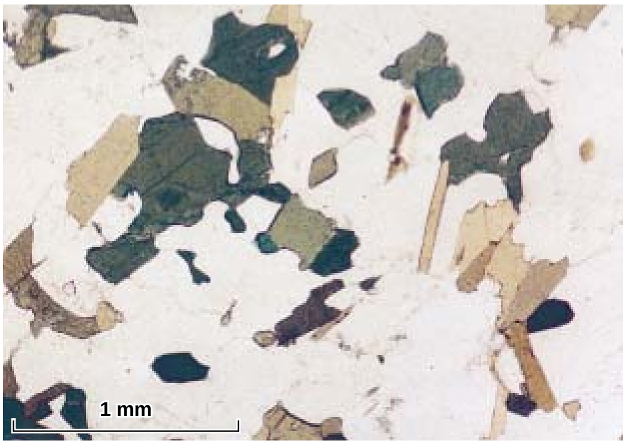

Fig. 34. Thin section of amphibolite facies hornblende-biotite orthogneiss; the ferromagnesian minerals are subhedral and in textural equilibrium, without signs of retrogression. GGU 289276, east coast of Qugssuk.

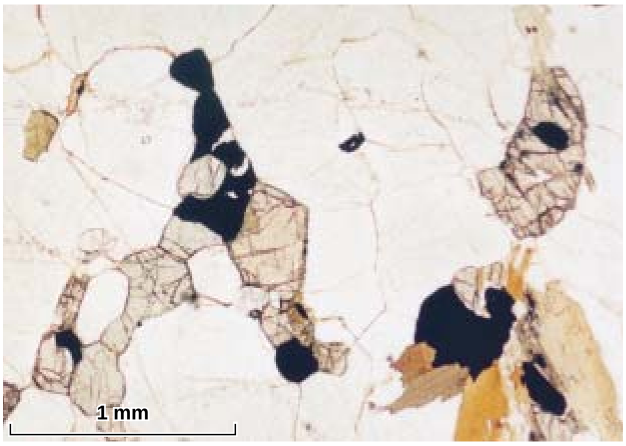

Fig. 35. Partially retrogressed granulite facies orthogneiss; ortho- and clinopyroxene are variably replaced by amphibole and biotite. GGU 278787, 8 km east of Ulamertoq.

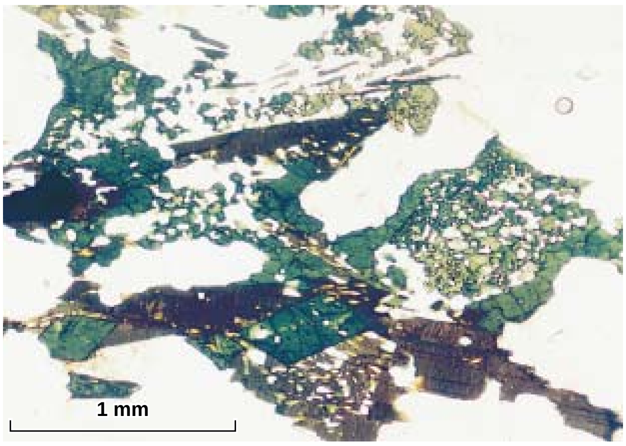

Fig. 36. Complete 'high-grade' retrogression in GGU 278787 (same thin section as in Fig. 35) with secondary, both euhedral and sieve-textured hornblende, and secondary biotite which in part also has sieve texture.

Tonalitic gneiss

The tonalitic orthogneiss which constitutes most of the Fiskefjord area typically forms tens of metres to a couple of kilometres thick composite sheets and layers in earlier supracrustal rocks and dioritic orthogneiss. Tonalitic orthogneiss in amphibolite facies (not retrogressed from granulite facies) at the east coast of Qugssuk has yielded a Rb-Sr whole-rock errorchron of 2954 ± 120 Ma and the above mentioned composite Pb-Pb whole-rock age of 3112 ± 40 Ma (see Garde, 1989a and pp. 79-81 for details). The tonalitic orthogneiss is broadly contemporaneous with the Taserssuaq tonalite complex, from which a conventional U-Pb zircon age of 2982 ± 7 Ma was reported by Garde *et al.* (1986; Table 1).

The tonalitic gneiss comprises numerous members of variable size and composition, which were successively emplaced into each other; much of the orthogneiss is polyphase even on outcrop scale. This is most easily recognised in the eastern amphibolite facies part of the area, where original boundaries between successive intrusive phases with slightly different colour indices are sharp and well preserved even where the rocks are strongly deformed (Fig. 31). Elsewhere the extensive recrystallisation accompanying first granulite facies and then retrogressive metamorphism has led to the development of blebby texture and more or less effectively destroyed such earlier lithological details (see Figs 32, 33; Garde, 1990, and discussion of blebby texture p. 81).

The tonalitic orthogneiss is generally medium-grained and composed of quartz, plagioclase (around An_{25}, commonly antiperthitic), minor interstitial K-feldspar, hornblende brown biotite or both, and accessory apatite, zircon, and magnetite-ilmenite intergrowths or pyrrhotite (Fig. 34). Granulite facies equivalents commonly contain both hornblende and hypersthene. In retrogressed gneiss hypersthene is variably replaced by fine-grained intergrowths of blue-green amphibole and quartz, and sheaves of secondary pale brown or green biotite commonly overgrow the iron-titanium oxides; large variations in the completeness of retrogression may be observed even within one thin section (Figs 35, 36). Occasionally two morphologically and compositionally different phases of biotite have been observed in retrogressed grey gneiss, typically in the form of sheaves of brown biotite a few millimetres large and much smaller interstitial flakes of greenish biotite with very low titanium contents (see section on geochemistry). The presence of two different biotite generations may suggest that more than one episode of retrogression occurred in parts of the area.

A distinctive field type of homogeneous tonalitic gneiss, characterised by relatively coarse grain size and conspicuous clots of retrograde amphibole and biotite, occurs on Qeqertaussaq and west of Serquartup imâ around central Fiskefjord. This rock type forms a roughly equidimensional body more than 5 km in diameter, although its exact boundaries have not been delineated and are partly concealed by the fjord. In terms of texture and general field appearance it resembles retrogressed, relatively mafic parts of the Taserssuaq tonalite complex described below, and like this complex it may have been emplaced late amongst the tonalitic orthogneisses. Another variety of partially retrogressed grey gneiss, mainly of granodioritic composition, occurs around the head of Fiskefjord and north of Tovqussap nunâ (Plate 1; Garde, 1989b). This variety has characteristic purplish weathering colours and was described as 'purple gneiss' by Berthelsen (1960). It is not certain if the purple colour is related to a particular composition, a particular stage of retrogression, or both.

The tonalitic orthogneiss commonly contains metre-sized enclaves of its host rocks, especially within a few hundred metres from the margins of larger bodies of supracrustal rocks. This spatial association suggests that most enclaves were only transported over short distances within the tonalite magmas. The most common enclaves are composed of amphibolite, but in the vicinities of the previously described layered complexes and around Ulamertoq both amphibolite and ultrabasic or ultramafic enclaves occur in the gneiss. Both dioritic and tonalitic orthogneisses in the south-western part of the Fiskefjord area, and on northern Tovqussap nunâ, contain numerous enclaves of leucogabbro and anorthosite. South of Igdlutalik, an inlet of outer Fiskefjord, there are very common elongate diorite enclaves in the tonalitic host, probably derived from neighbouring dioritic bodies.

Trondhjemitic gneiss

Certain areas up to a few square kilometres are dominated by trondhjemitic, biotite-bearing orthogneiss. However, in most places this lithology could not be clearly differentiated in the field from the main phase of tonalites. One such area is located west of the head of Fiskefjord, where trondhjemite sheets up to *c.* 500 m thick were emplaced into supracrustal amphibolite and subsequently participated in repeated folding. Other, similarly deformed sheets occur at the outer coast south of Fiskefjord. Like the adjacent tonalitic gneiss these trondhjemites display distinct recrystallised, blebby textures although the mafic, mainly biotitic, clots are few and far apart. Trondhjemite emplacement, deformation and metamorphism matches that of the surrounding tonalitic gneiss, and the trondhjemites are believed to be cogenetic with the tonalitic gneiss. Other trondhjemites form the cores of late, elongate dome-like structures up to a few kilometres in size, for example north of Qôrngualik south of outer Fiskefjord.

Fig. 37. Flat-lying sheets of synkinematic granitic gneiss (mesoperthite granite) with discordant margins to dioritic gneiss. 1 km south of Quagssûp alánguata tasê in the central part of the Fiskefjord area.

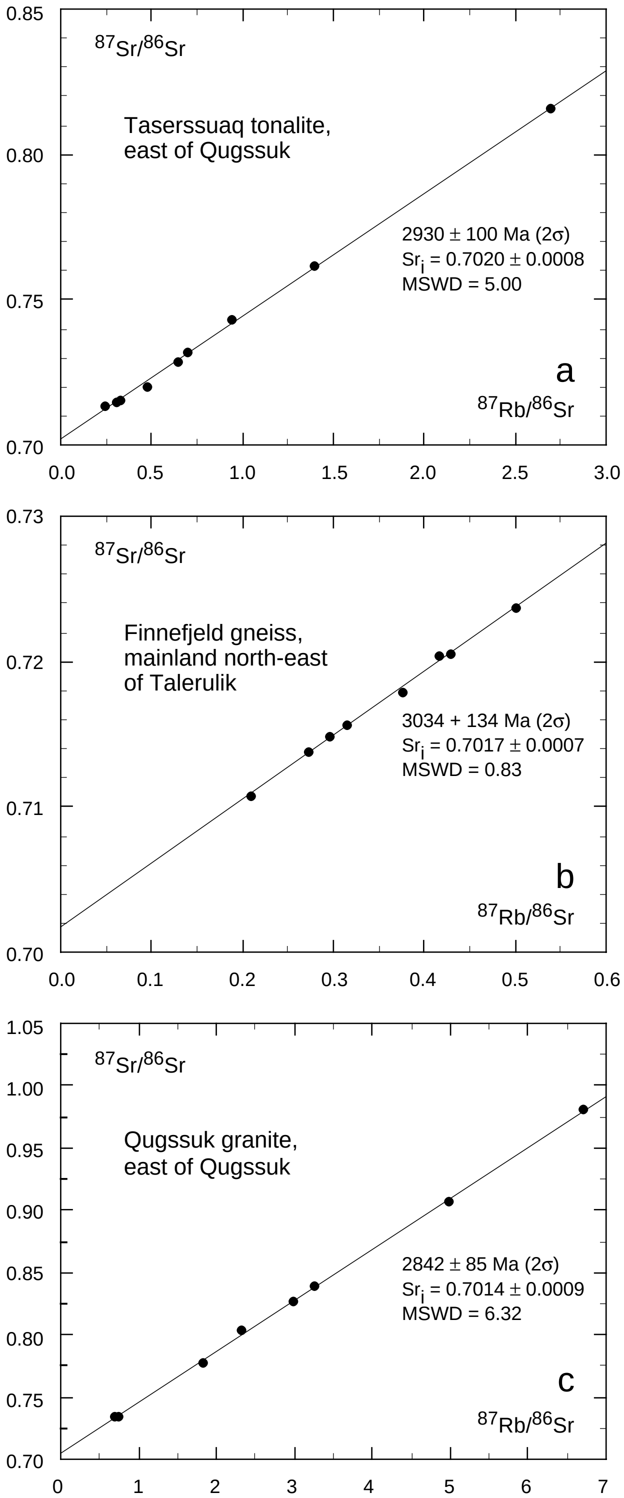

Fig. 38. Isochron diagrams of not previously published Rb-Sr data. (**a**) Taserssuaq tonalite east of Qugssuk, (**b**) Finnefjeld gneiss, and (**c**) Qugssuk granite east of Qugssuk (See also Table 1). The data set is listed in Table 2.

Trondhjemitic rocks also occur within the granulite facies dioritic gneiss in the south-western part of the Fiskefjord area, usually as thin sheets. Biotite is normally the only mafic mineral, but dark antiperthitic plagioclase and smoky blue quartz suggest pre- or syn-granulite facies emplacement. Similar rocks occur further south in Nordlandet, where they have been termed Kua granulite (Macdonald, 1974) and Imartuneq granite complex (Reed, 1980). Both the latter author and McGregor (1993) suggested that they are products of partial anatexis during granulite facies metamorphism.

Granodioritic and granitic gneiss

Small bodies of leucocratic granodiorite and granite, most of which have been emplaced before or during the granulite facies event, occur in the central and western parts of the Fiskefjord area. The granitic rocks most commonly form sheets of very variable thickness (from a few centimetres to maximum *c.* 1 km), which are locally discordant to gneissic foliation in their host rocks. They have frequently been emplaced more or less concordantly along boundaries between amphibolite and orthogneiss (e.g. on Tovqussap nunâ and south-west of Quagssûp alánguata tasê), as sheets subparallel to earlier foliation in dioritic gneiss (Fig. 37), or they form irregular bodies within tonalitic or dioritic gneiss (e.g. south of outer Fiskefjord). An elongate, *c.* 2 km long granite dome occurs at Oqúmiap taserssua. Around the sound Pâtôq north of Tovqussap nunâ granodioritic and granitic rocks constitute larger areas that may represent continuations of the largest sheets on northern Tovqussap nunâ. The granitic rocks are medium grained and biotite-bearing and have granular, equidimensional mineral textures suggesting more or less complete solid state recrystallisation. They either contain mesoperthite feldspar (indicative of crystallisation above the alkali feldspar solvus, i.e. under granulite facies conditions), or feldspar mosaics of antiperthitic plagioclase and microcline (probably also recrystallised from an intermediate alkali feldspar precursor). Biotite is concentrated in rare aggregates a few millimetres large, and magnetite is a common accessory.

Local granitic rocks within the granulite facies areas (e.g. south-west of Kingigtoq) post-date the peak metamorphic event and may be contemporaneous with syn- or post-retrogression granitic rocks in the eastern part of the Fiskefjord area (see pp. 45–48).

Taserssuaq tonalite complex

The Taserssuaq tonalite complex is a large, homogeneous, late-tectonic body akin to the grey gneiss. It was first identified during reconnaissance mapping (Allaart *et al.*, 1977) and was briefly described by Garde *et al.* (1983), Kalsbeek & Garde (1989) and Nutman & Garde (1989). The complex covers more than 1500 km^2 in the Isukasia area (Garde, 1987) and the north-eastern part of the Fiskefjord area. It forms an important element in the evolution of the Akia terrane, and a summary description is presented here. Garde *et al.* (1986) interpreted a zircon U-Pb age of 2982 ± 7 Ma (analyst R. T. Pidgeon, Western Australian Institute of Technology) as the age of intrusion. Rb-Sr data (Garde *et al.*, 1986) gave a whole rock age of 2882 ± 36 Ma (initial $^{87}Sr/^{86}Sr$ = 0.7017 ± 0.0002, MSWD = 1.57), and Rb-Sr mineral isotopic data indicated thermal events at *c.* 2500 and 1700 Ma. The Rb-Sr age was interpreted by Garde *et al.* (1986) as probably reflecting mobility of Rb associated with retrogression, as suggested by Kalsbeek & Pidgeon (1980) for orthogneisses in the Fiskenæsset area, southern West Greenland (see also pp. 79–81). A subset of samples from the Rb-Sr data set discussed above, collected from predominantly granodioritic rocks in the southernmost part of the complex east of Qugssuk, gives an age of 2930 ± 100 Ma (initial $^{87}Sr/^{86}Sr$ = 0.7020 ± 0.0008, MSWD = 5.00), see Table 2 and Fig. 38a. These samples do not appear to have been subject to retrogression.

The Taserssuaq tonalite is locally discordant to the host grey gneiss, although most boundaries are ambiguous or transitional, and its emplacement post-dates early grey gneiss – amphibolite folds. The western boundary zone possesses a steep but commonly indistinct schistosity developed during syn-emplacement deformation (Garde *et al.*, 1983). The interior of the complex generally has a weak flat-lying S fabric and is less deformed than its margins and the surrounding gneiss. Nutman & Garde (1989) presented evidence that the Taserssuaq tonalite complex is mushroom-shaped and was emplaced in a diapiric fashion.

Most of the complex consists of homogeneous tonalitic rocks; granodioritic and granitic members occur in addition to tonalite in its southern part. This part of the complex also contains lensoid enclaves and sheets of homogeneous or indistinctly layered, medium-grained, hornblende-bearing dioritic and gabbroic rocks, which vary in length from decimetres up to *c.* 2 km. Their chemical compositions (see p. 65) suggest that most of them are cogenetic with the tonalite complex; the absence of such rocks in the neighbouring grey gneiss supports this interpretation.

A shallowly dipping compositional layering of igneous origin can locally be seen within the central part of the complex. The layering is defined by modal variations of hornblende and plagioclase and may stretch over hundreds of metres. Such layering, which is not usually found within sheeted grey gneiss of similar composition, indicates that individual batches of Taserssuaq tonalite magma had more time to differentiate during crystallisation and were therefore larger than the batches of magma that formed the earlier sheets of grey gneiss. In addition to local igneous layering the complex is characterised by subhedral, 1–2 cm large feldspar crystals set in a medium-grained quartzo-feldspathic matrix with granitic texture, and scattered, few millimetres large hornblende-biotite aggregates. A small part of the complex east of lake Taserssuaq carries granulite facies parageneses (Allaart, 1982), and some areas within the central and particularly the western parts of the com-

Table 2. Rb-Sr whole rock data for Finnefjeld gneiss complex, Taserssuaq tonalite complex and Qugssuk granite

	Rb ppm	Sr ppm	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
Finnefjeld gneiss complex				
339638	47	323	0.415	0.7203
339644	27	381	0.209	0.7108
339645	44	459	0.272	0.7138
339647	53	354	0.430	0.7205
339648	56	323	0.500	0.7237
339650	38	360	0.315	0.7156
339652	53	414	0.376	0.7179
339655	48	466	0.295	0.7148
Taserssuaq tonalite complex				
283315	79	488	0.479	0.7200
283319	128	139	2.693	0.8157
283321	90	281	0.944	0.7433
283328	80	328	0.692	0.7318
283334	76	163	1.399	0.7612
283340	72	325	0.640	0.7286
283341	55	506	0.327	0.7155
283372	45	540	0.245	0.7131
283373	60	572	0.304	0.7145
Qugssuk granite				
283322	104	104	2.974	0.8258
283350	126	159	2.316	0.8031
283367	105	414	0.736	0.7335
283375	167	268	1.815	0.7765
283376	87	358	0.701	0.7339
283377	142	85	4.973	0.9064
283378	133	60	6.693	0.9800
283379	147	132	3.259	0.8393

Analytical methods as described by Garde *et al.* (1986); the precision of Rb/Sr measurements is within *c.* 1% (2σ), and of $^{87}Sr/^{86}Sr$ measurements better than 0.0002 (2σ). See also Fig. 38.

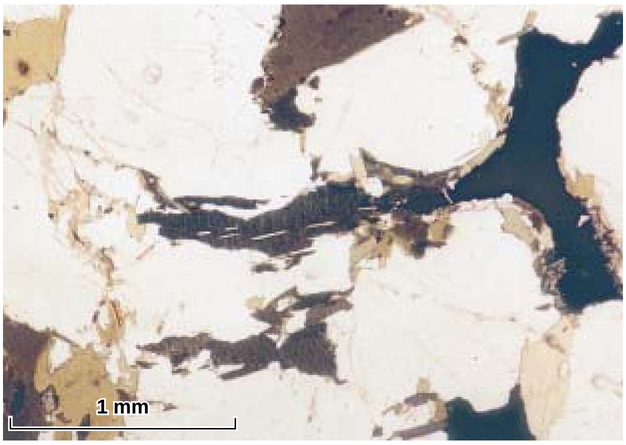

Fig. 39. Thin section of biotite-bearing Taserssuaq tonalite in amphibolite facies. Small irregular biotite grains along the margins of larger biotite grains indicate partial retrogression of the primary amphibolite facies paragenesis. GGU 289287, 7 km north-east of the head of Qugssuk.

plex commonly have recrystallised, blebby textures indicative of retrogression from granulite facies. Growth of secondary biotite was also observed in some rocks with primary amphibolite facies parageneses (Fig. 39), which suggests that the effects of retrogression extended beyond the area formerly of granulite facies.

In summary, the Taserssuaq tonalite complex is in many ways similar to the surrounding grey gneiss but was intruded relatively late and in larger batches. Inclusions of supracrustal rocks are uncommon, but there are cogenetic dioritic and gabbroic enclaves.

Finnefjeld gneiss complex

The Finnefjeld complex was first described by Berthelsen (1957, 1962), who studied its north-western boundary area. The main part of the complex is located in the Maniitsoq (Sukkertoppen) district to the north, but its south-eastern margin reaches down into the Fiskefjord area (Plate 1). No precise age has so far been obtained from the Finnefjeld gneiss complex; the field evidence seems to suggest that it may be younger than the Taserssuaq tonalite complex. A conventional U-Pb zircon age of 3067 $^{+62}_{-42}$ Ma was obtained from a sample collected 3 km south-east of Sisak (GGU 339641: one concordant and two discordant size fractions of light brown, idiomorphic to slightly corroded crystals, B. T. Hansen, personal communication, 1990). A Rb-Sr whole-rock age of 3058 ± 123 Ma (initial $^{87}Sr/^{86}Sr$ = 0.7012 ± 0.0007, MSWD = 4.7), obtained by S. Moorbath, University of Oxford, from samples scattered along the mainland coast, was reported by Garde (1990). A second Rb-Sr data set from samples collected around Sisak in the southern part of the complex is very similar (3034 ± 140 Ma, initial $^{87}Sr/^{86}Sr$ = 0.7017 ± 0.0007, MSWD = 0.83), see Fig. 38b and Table 2. Pb-Pb whole rock analysis of 12 samples collected on the island Talerulik gave a best fit line of 2700 $^{+380}_{-350}$ Ma, MSWD = 20.74, model $\mu1$ = 7.59 (P. N. Taylor, personal communication, 1987); this data set is difficult to interpret but may suggest an isotopically inhomogeneous source, contamination with unradiogenic Early Archaean lead during magma emplacement, or later Pb metasomatism, or both.

The interior parts of the Finnefjeld gneiss complex are only poorly known from helicopter reconnaissance (Allaart *et al.*, 1978) and from unpublished surveys during mineral exploration by Kryolitselskabet Øresund A/S. Its southern boundary zone on the mainland north of Tovqussap nunâ and adjacent islands along the northern boundary of the Fiskefjord map area were surveyed in detail by Marker & Garde (1988) who revised earlier interpretations of the complex.

The Finnefjeld gneiss complex resembles the Taserssuaq tonalite complex in several ways. Both complexes consist of a number of large, generally very homogeneous intrusions of predominantly tonalitic rocks which locally preserve igneous layering, and which were emplaced into already folded and metamorphosed units of supracrustal rocks and Middle Archaean grey gneiss. The main difference between the two complexes is that the Finnefjeld gneiss clearly post-dates the granulite facies event. It has indisputable intrusive contacts with grey gneiss and supracrustal rocks, and in the south-eastern border zone transects their common large-scale structures at large angles (Plate 1; Marker & Garde, 1988, fig. 2). Four intrusive phases were identified, all of which have equilibrium amphibolite facies parageneses without signs of earlier granulite facies metamorphism or pervasive retrogression; Fig. 40 shows a typical lithology. Marker & Garde (1988) noted, however, that the complex also contains local older gneissic enclaves with a more complex history. The latest intrusive phases of Finnefjeld gneiss are granodioritic to granitic in composition and constitute a hybrid border zone along the grey gneiss, which probably includes partially remobilised material derived from the latter rocks.

Whereas the south-eastern hybrid border zone itself is an area of low strain, several NNE- to NE-trending zones of strong deformation occur inside the complex,

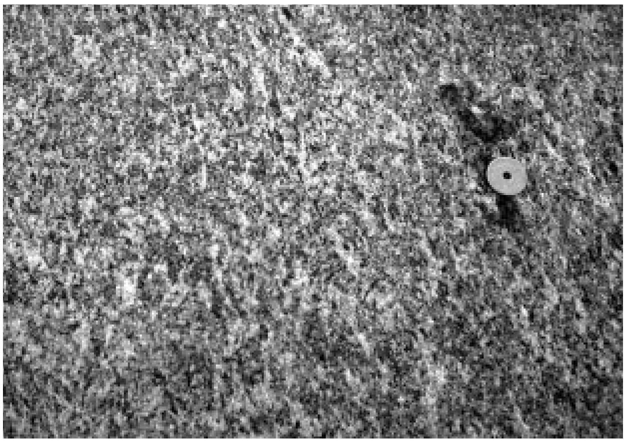

Fig. 40. Typical exposure of homogeneous, medium-grained Finnefjeld gneiss with elongate plagioclase aggregates set in a granitic matrix. 500 m north of Sisak.

e.g. on the island Talerulik and in the mainland north of Sisak (Fig. 41). These high strain zones have a distinct steep to vertical schistosity, commonly combined with tight upright small-scale and larger folds and a shallow south to south-westerly plunging mineral or intersection lineation parallel to fold axes. Marker & Garde (1998) tentatively concluded that these structures were developed in a NW–SE compressional stress field during or soon after intrusion in an overall transpressional setting. A component of transcurrent simple shear may also have been present.

Fig. 41. NNE-trending high strain zone in Finnefjeld gneiss about 1.5 km north of Sisak with tight upright folds and axial planar pegmatite.

Emplacement mechanisms of tonalitic magma

Nutman & Garde (1989) discussed emplacement mechanisms of grey gneiss and Taserssuaq tonalite. They noted that the supracrustal rocks were intruded first by sheeted grey gneisses and then by the domal Taserssuaq tonalite complex. At early stages of magmatic accretion, the tonalitic magma would be emplaced into shallow, cool supracrustal rocks with up to about 5 wt% water, and would exploit hydraulic fractures. If the fractures were formed in an environment of extension or moderate compression, the intrusive sheets would be shallow or moderately inclined, assuming subhorizontal orientation of the maximum stress. Subsequent larger batches of magma such as those forming the dome-shaped Taserssuaq tonalite or Finnefjeld gneiss complexes would be intruded into sialic crust that was already thickened, drier (*c.* 1 wt% water), but now also much hotter and more ductile.

Structural evolution

A general picture of the deformational history that accompanied magmatic accretion of middle continental crust in the Fiskefjord area – sheets of grey gneiss, large tonalite complexes, and late granites – can be acquired by integrating observations in different parts of the area. Around central and outer Fiskefjord interleaved supracrustal rocks and grey orthogneiss outline well-exposed interference patterns between two or three sets of superimposed folds. In other parts of the area interference patterns have been modified by prominent younger, N–S trending high strain zones, and by domes with cores of younger granites. Other parts of the area, those underlain by the Taserssuaq tonalite and Finnefjeld gneiss complexes, have not been subject to strong penetrative deformation.

Interpretation is complicated because most structures in the Fiskefjord area are products of both semi-

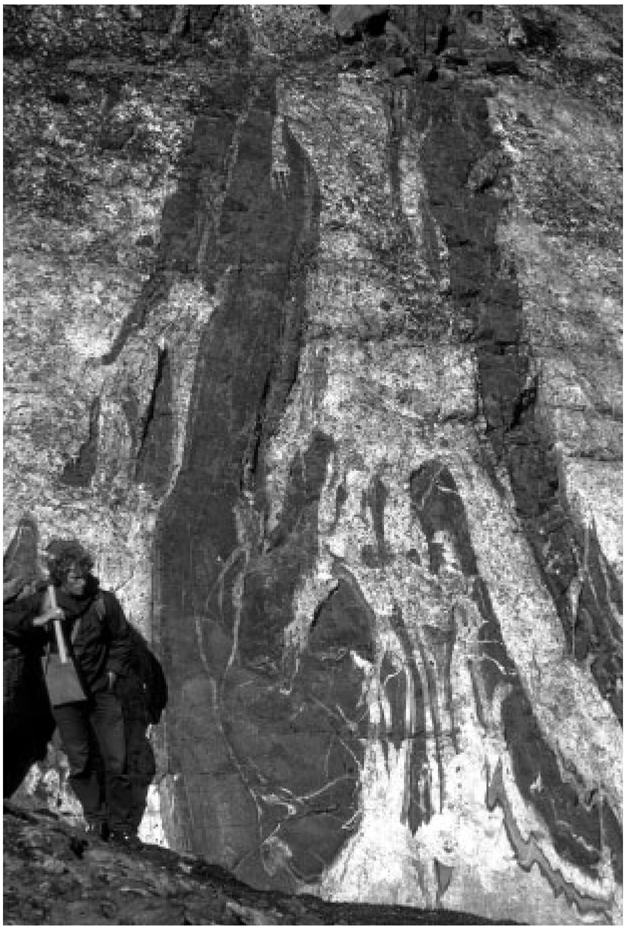

Fig. 42. Complex relationship between supracrustal amphibolite and polyphase orthogneiss, produced by both igneous processes (the splitting and interfingering nature of the intrusive leucocratic rock) and solid-state ductile deformation (isoclinally folded amphibolite screen and *s*-shaped small-folds in right part of the picture). 8 km east of Ulamertoq.

crystalline deformation during magma emplacement and subsequent solid state deformation, as can commonly be observed on outcrop scale (Fig. 42). Besides, it has been demonstrated both in the field and experimentally that several successive events of deformation are not always necessary to produce superimposed folds. For example, Ramberg (1967) and Wikström (1984) have shown by experiments and field studies that 'synchronous refolding' can occur, for instance, by diapiric movements in ductile rocks, or by interaction between gravity-induced diapiric structures and contemporaneous folding in response to a lateral stress field. However, for reasons given below it is plausible that at least three separate phases of folding involving grey gneiss have occurred in the Fiskefjord area.

Distribution of supracrustal rocks within the orthogneiss

Much of the deformation history of the Fiskefjord area is monitored by the distribution and shapes of bodies of amphibolite and related supracrustal rocks, which form structural markers in the otherwise rather monotonous orthogneiss. These patterns also reflect the primary relationships between the two groups of rocks as discussed elsewhere.

In most of the Fiskefjord area, namely where tonalitic and trondhjemitic orthogneisses predominate, older supracrustal rocks form up to about 20 per cent of the outcrop (Plate 1). In some of these areas, for instance north of Qugssuk and on Tovqussap nunâ, they form many kilometres long, multiply folded layers which are linked with each other through complex fold structures that also incorporate the intercalated intrusive orthogneiss. In other areas such as north of outer Fiskefjord, amphibolite and related rocks mostly occur as separate enclaves ranging between a few centimetres and several kilometres in length.

Supracrustal rocks are less common in the large tracts of dioritic orthogneiss that constitute most of the mainland south of Fiskefjord and Nordlandet farther south. The Taserssuaq tonalite and the Finnefjeld gneiss complexes appear to be largely without such enclaves, although their margins have in some places intruded and agmatised supracrustal units.

Outline of principal structural elements

The structural elements in various parts of the Fiskefjord area that together mark its structural evolution may well be representative of the entire Akia terrane. The north-western, north-eastern and south-eastern parts of the area are all characterised by complex, multiply folded outcrop patterns, frequent changes in the orientations of structural elements, and structures up to about 10 km in size (Berthelsen, 1960; Lauerma, 1964; Garde, 1986; Garde *et al.*, 1987). These structures appear to have been developed during the early phases of the *c.* 3000 Ma continental crustal accretion. The central and southern parts are dominated by somewhat larger structures in the lithologically very homogeneous, granulite facies areas of *c.* 3220 Ma Nordlandet dioritic gneiss. The dioritic rocks contain sporadic trains of small leucogabbroic and other inclusions, which outline large recumbent folds, e.g. north of Oqúmiap taserssua. As in the above mentioned areas these recumbent folds are folded by smaller upright folds (see p. 41).

Several other tracts are characterised by prominent, N–S trending structure with distinct steeply dipping schistosity, subhorizontal linear elements and large isoclinal folds. Some of these N–S trending high strain zones occur along the margins of dioritic gneiss units in the southern part of the area. Another, major high strain zone follows the west coast of inner Godthåbsfjord and continues northwards through the head of Fiskefjord; this zone was established before or during the granulite facies event, with later reactivation. Still others occur in the peninsula east of Qugssuk, where they separate areas of open and variable structure. The Finnefjeld gneiss complex contains several NNE-trending high strain zones close to its south-eastern margin which may possibly be related to the above structures although they are younger (see p. 45).

Several dome-shaped structures were developed late in the structural evolution; most have cores of leucocratic tonalite, trondhjemite or granite. Lauerma (1964) described the almost 10 km large Ipernat dome west of Nãngissat, and Berthelsen (1950) mapped a smaller one around Qaersup ilua in the western part of Tovqussap nunâ. Other domes or partial domes with cores of leucocratic orthogneiss occur at Qeqertaussaq, south of Oqúmiap taserssua, at Quagsugtarssuaq, and north of Narssarssuaq. The Igánánguit granodiorite (see p. 45) forms a *c.* 20 km long, NNE-trending composite dome *c.* 10 km north-east of the head of Fiskefjord.

Early phases of deformation on Tovqussap nunâ

Berthelsen (1960) published a detailed account of the structural evolution of Tovqussap nunâ in the northwestern part of the Fiskefjord area and recognised four main phases of deformation. The first two, the Midterhøj and Smalledal phases, resulted in large recumbent isoclinal folds with NW-, and ENE- to NE-trending axes respectively. The Tovqussap dome, a prominent structure in the western part of Tovqussap nunâ, was interpreted to have formed by a combination of several factors, including both antiformal folding of earlier structures and (?diapiric) movement of material towards the top of the structure. Berthelsen placed the dome-forming episode after the two first phases of deformation. The subsequent Pâkitsoq phase resulted in a series of conspicuous upright to overturned, open to tight folds of moderate size with SE- to S-plunging axes. Their wavelengths are *c.* 2–3 km (much shorter than those of the recumbent folds), their hinge zones are more angular, and they were commonly accompanied by a new axial planar foliation and mineral or rodding lineation. The Pâkitsoq phase of deformation took place contemporaneously with the culmination of metamorphism under granulite facies conditions. Berthelsen (1960) did not discuss the possibility of 'synchronous refolding', but the above mentioned differences between the Smalledal phase of recumbent folds and the Pâkitsoq phase of upright folds suggest that this possibility is unlikely.

Berthelsen's structural analysis led him to assume that Tovqussap nunâ originally consisted of a conformable, around one kilometre thick pile of supracrustal rocks. This was a reasonable conclusion at the time, considering the prevailing conceptions about the origin of grey gneiss; it was not widely recognised until a decade later that the bulk of Archaean grey gneisses are of intrusive igneous origin (e.g. McGregor, 1973). The granitic rocks on Tovqussap nunâ were interpreted by Berthelsen (1960) as formed by *in situ* granitisation, not as (late) intrusive bodies. Berthelsen (1960, p. 212) was aware that part of his structural analysis would fall apart if some or all of the isoclinal fold closures he had described or interpreted could instead be shown to be the tips of wedge-shaped intrusions. However, the geological map (Plate 1 in Berthelsen, 1960) that forms the basis for his analysis is very accurate and mostly covers well-exposed ground. Accordingly, a substantial part of the structural evolution he proposed for Tovqussap nunâ can be recognised with appropriate modifications in much of the Fiskefjord area.

One element that warrants revision or at least commenting is Berthelsen's (1960, p. 147) earliest, Midterhøj phase of deformation. Berthelsen based the Midterhøj phase on an interpretation of one apparent isoclinal fold closure outlined by an amphibolite with a core of leucocratic gneiss, and another inferred, unexposed closure, both located in the upper flank of a later fold. The Midterhøj phase was only recognised in the eastern part of Tovqussap nunâ, and in view of the intrusive origin of the rock in the core of Berthelsen's supposed isoclinal closure, it is unlikely that the Midterhøj phase exists here. However, an early phase of deformation with isoclinal folds of similar orientation has later been documented from central Fiskefjord. These folds were refolded during two events which match Berthelsen's Smalledal and Pâkitsoq phases (see the following sections).

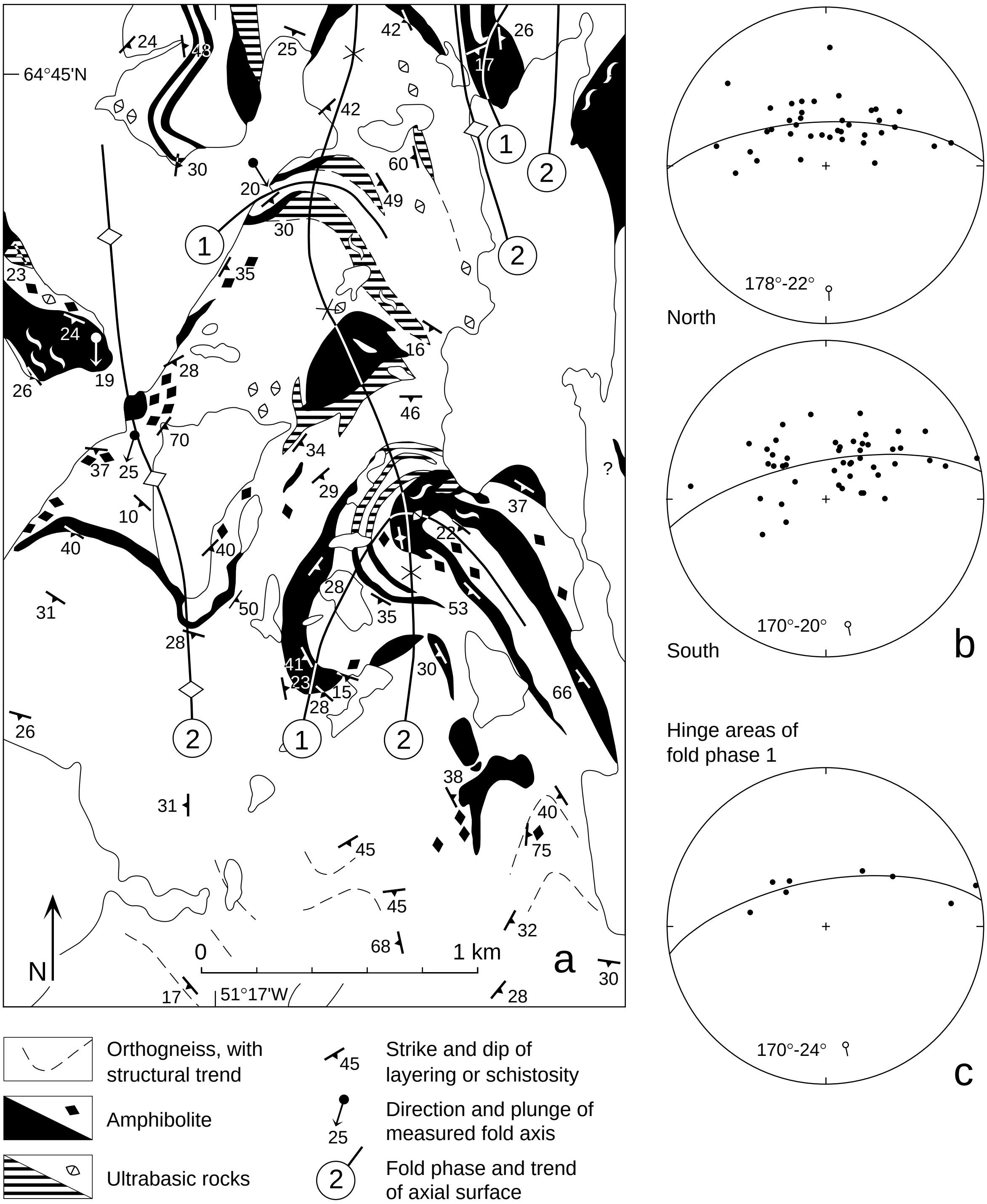

Fig. 43. (**a**) Superimposed folding north-east of Itivneraq, south of outer Fiskefjord. Axial surfaces of N–S trending folds belonging to the Påkitsoq phase are marked as (2) and have shallow south-plunging axes. These folds are superimposed on an earlier set of overturned to recumbent isoclinal folds (1), which also have south-plunging axes but originally had E–W trending axial surfaces.
(**b**) Stereographic projections of poles to layering and schistosity from the northern and southern parts of the area. In both subareas the poles show considerable scatter relative to a constructed great circle but indicate a persistent southerly axial plunge.
(**c**) Strike and dip measurements of layering and schistosity within two hinge zones of the early folds (phase 1) suggest that also the early folds had southerly axial plunges.

Thrusting of orthogneiss and amphibolite on Angmagssiviup nunâ prior to isoclinal folding

In the western part of Angmagssiviup nunâ, M. Marker (personal communications, 1987, 1995) mapped an up to *c.* 200 m thick, compositionally banded amphibolite – grey gneiss sequence (Garde, 1989b). The banded sequence consists of tonalitic gneiss in granulite facies and retrogressed to amphibolite facies with closely spaced, parallel bands of amphibolite on a scale of centimetres to decimetres, and the composite rock has distinct schistosity. The banded sequence can be followed through isoclinal fold closures belonging to the earliest phase of folding (see below) and is interpreted by M. Marker as having resulted from tectonic mixing of orthogneiss and amphibolite units during early thrusting at a shallow crustal level. The banded sequence is thus considered to represent the earliest episode of deformation that comprises both the supracrustal association and the *c.* 3000 Ma grey gneiss.

Refolded folds around central and outer Fiskefjord

Garde *et al.* (1987) described two areas between Igdlutalik and Tasiussarssuaq north of central Fiskefjord, and south-west of Serquartup imâ south of central Fiskefjord, where early isoclinal folds with amplitudes of several kilometres are folded by a second set of large, flat-lying, E–W trending isoclinal folds. These are in turn refolded by upright, NE–SW trending open folds with smaller wavelengths (in the order of 1–2 km) and shallow NE-trending doubly plunging axes. Garde *et al.* (1987) ascribed these folds to three successive phases of deformation, which closely correspond to the Midterhøj, Smalledal and Pâkitsoq phases of Berthelsen (1960).

Other early isoclinal folds occur, for instance, in granulite facies Nordlandet dioritic gneiss west of Qôrngualik; it is possible that the earliest structures in this region predate the *c.* 3000 Ma crustal event. McGregor (1993, fig. 5) described a large synclinorium with a width of *c.* 15 km south-west of Natsigdlip tasia (in the border area between the Qôrqut and Fiskefjord map areas), the northern part of which refolds an isoclinal fold outlined by trains of leucogabbroic inclusions west of Qôrngualik. The synclinorium is refolded by a series of asymmetric south-plunging folds with wavelengths less than 1 km, which are contemporaneous with the *c.* 3000 Ma granulite facies metamorphism and can be correlated with folds of the Pâkitsoq phase further north.

The Pâkitsoq phase of upright to overturned, NE–SW to N–S trending folds with wavelengths less than *c.* 5 km is recognised in most parts of the Fiskefjord area. Figure 43 shows an example east of Itivneraq, outer Fiskefjord. Here, folds of the Pâkitsoq phase with south-plunging axes refold recumbent folds with similar fold axis orientations. Garde *et al.* (1987) described NNW-trending structures with apparent domes and basins in the north-western part of the Fiskefjord area, e.g. north-west of Tasiussarssuaq. These are now interpreted as folds belonging to the Pâkitsoq phase with axial culminations and depressions controlled by pre-existing structures.

North–south trending high strain zones

Several localised N–S and NNE-trending, steep to vertical zones of relatively late high strain occur in the central and eastern parts of the Fiskefjord area. The most prominent is the Qugssuk–Ulamertoq zone (Fig. 44). This runs along the west coast of Qugssuk, continues northwards towards Fiskefjord along the eastern shoulder of the ridge Ulamertoq, and from there follows the head of Fiskefjord further north into the Isukasia map area (with a dextral offset of *c.* 2 km along the Proterozoic Fiskefjord fault, see p. 85). The zone largely comprises orthogneiss of tonalitic and dioritic composition, several large amphibolites, and a prominent, *c.* 25 km long and up to *c.* 800 m wide layer of dioritic gneiss which crosses Fiskefjord. Some of the thickest amphibolite bodies in the Qugssuk–Ulamertoq zone outline large upright isoclinal folds with steep to vertical limbs. Their fold axes plunge *c.* 15° S, and fold axis-parallel hornblende lineation is common. In the hinge zones small scale, recumbent, shallow south-plunging isoclinal folds are outlined by amphibolite interleaved with thin orthogneiss layers. The latter folds are believed to predate the upright isoclinal folds. Besides, small, more or less symmetrical isoclinal folds with subvertical axial surfaces occur in the limbs of the regional folds; these folds could be either reorientated earlier folds or parasitic to the regional folds.

A large part of the orthogneiss within the zone is retrogressed and has an irregular quartzo-feldspathic netveining and a diffuse, undeformed, blebby texture of secondary, centimetre-sized biotite-amphibole patches (e.g. Fig. 32). Areas of granulite facies tonalitic gneiss

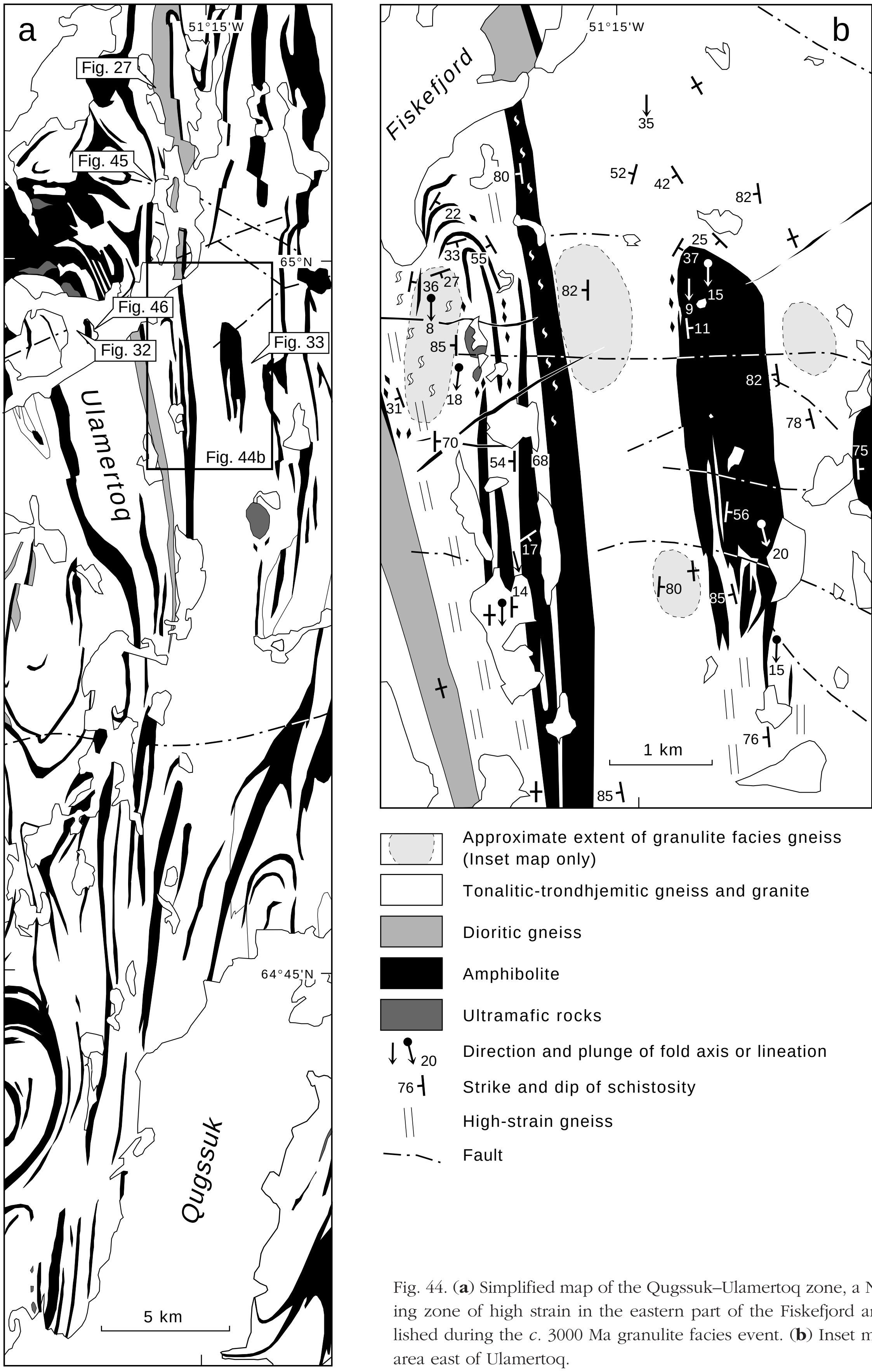

Fig. 44. (**a**) Simplified map of the Qugssuk–Ulamertoq zone, a N–S trending zone of high strain in the eastern part of the Fiskefjord area established during the *c.* 3000 Ma granulite facies event. (**b**) Inset map of the area east of Ulamertoq.

Fig. 45. Partially retrogressed grey gneiss with indistinct quartzo-feldspathic network. Qugssuk–Ulamertoq zone 200 m west of the head of Fiskefjord.

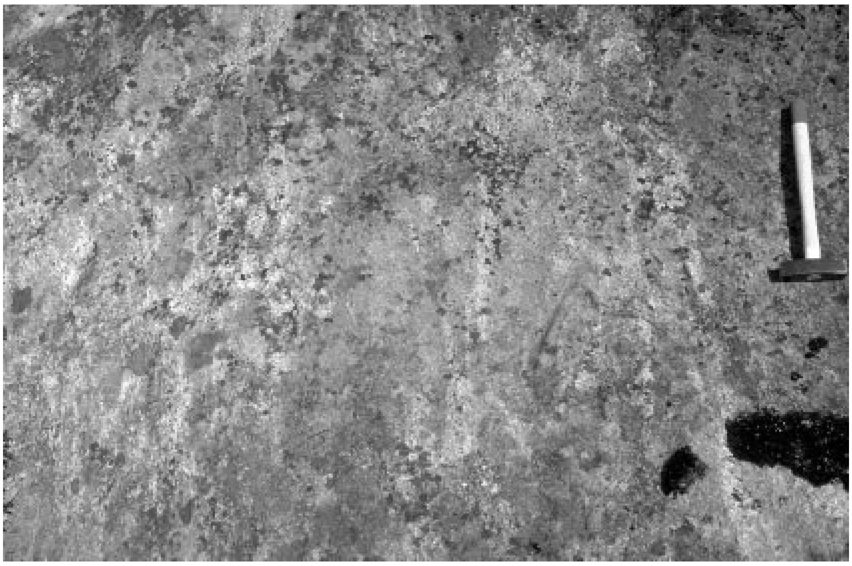

Fig. 46. Retrogressed orthogneiss with folded enclave of amphibolite. The strong planar fabric of the retrogressed orthogneiss in the central part of the picture suggests syn-retrogression deformation in this zone. East coast of island in inner Fiskefjord *c.* 5 km east of Kûlik.

with equigranular mineral textures, which have partially or completely escaped retrogression, cover irregular areas up to several square kilometres in size inside the Qugssuk–Ulamertoq zone (Fig. 44). Also, some parts of the dioritic gneiss layer that crosses the inner part of Fiskefjord have retained granulite facies parageneses and may have a distinct (granulite facies) schistosity (Fig. 27), whereas other parts (e.g. just west and south of Fiskefjord) have secondary amphibolite facies textures developed during static retrogression (Fig. 45).

The Qugssuk–Ulamertoq zone also contains local narrow tracts of retrogressed orthogneiss, a few decimetres to tens or hundreds of metres wide, with a distinct late schistosity that affects the retrograde biotite-amphibole patches (Fig. 46). In these tracts the thin quartzo-feldspathic veins commonly visible in retrogressed gneiss are reorientated and parallel to the new schistosity. Likewise, observations by V. R. McGregor (personal communication, 1995) along the western coasts of Godthåbsfjord and Qugssuk indicate that ductile deformation proceeded there after the onset of retrogression.

The relationships described above between map- and outcrop-scale structures and granulite facies and retrogressive textures in the Qugssuk–Ulamertoq zone indicate that its pervasive N–S structural grain, including the apparent transposition of earlier folds into upright isoclines, was developed while granulite facies conditions still prevailed.

The Qugssuk–Ulamertoq zone and related N–S trending structures are believed to have formed principally

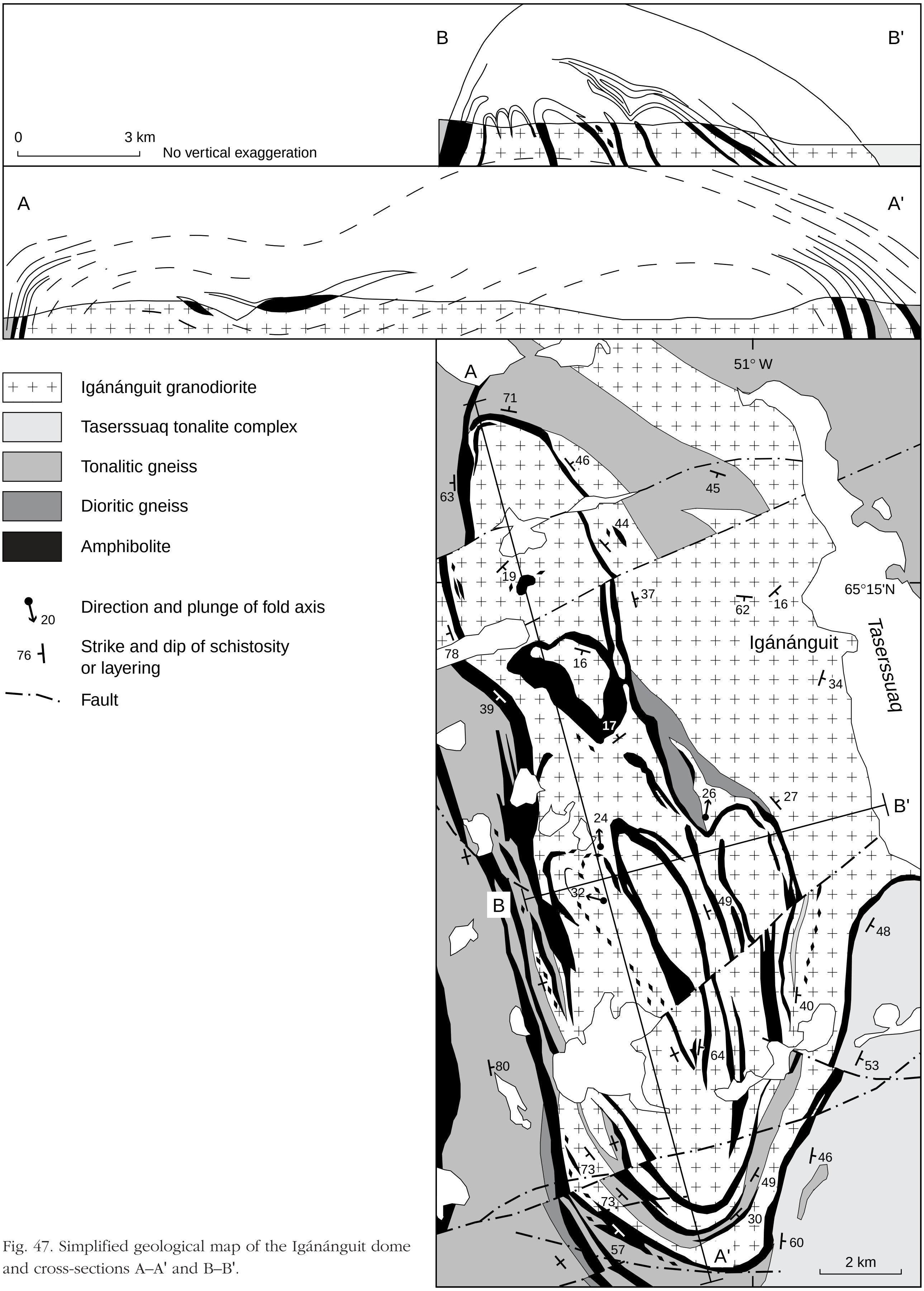

Fig. 47. Simplified geological map of the Igánánguit dome and cross-sections A–A' and B–B'.

in response to E–W compressional forces. Subhorizontal elongation is indicated by the common mineral lineation and shallow south-plunging orientations of fold axes. It is possible that this elongation was also related to a transcurrent component, but no direct evidence for this has been observed. The previously mentioned NNE-trending zone of strong deformation in the south-eastern part of the Finnefjeld gneiss complex is possibly related to the N–S trending high strain zones. Its trend is a little different and it post-dates granulite facies metamorphism, but it may overlap in time with late movements of the N–S zones and has a similar element of subhorizontal stretching.

McGregor *et al.* (1991) and McGregor (1993; personal communication, 1995) suggested that the high-strain zone along the western part of outer Godthåbsfjord and Nuuk town is continuous with the Qugssuk–Ulamertoq zone, and that both developed after the assembly of the Akia and Akulleq terranes at around 2700 Ma. While the Qugssuk–Ulamertoq zone appears to have been reactivated in connection with the terrane assembly, the available field evidence clearly suggests that it came into existence at about 3000 Ma under granulite facies metamorphic conditions.

Late granitic domes and sheets

Garde *et al.* (1986) described the Igánánguit granodiorite (formerly Igánánguit pink gneiss) and Qugssuk granite, two younger granitoids that crop out in the eastern part of the Fiskefjord area (age data follow below). Coastal outcrops of granitic rocks belonging to the Qugssuk granite were already shown on the map by Noe-Nygaard & Ramberg (1961). The Igánánguit granodiorite is confined to a composite dome west of the lake Taserssuaq. The Qugssuk granite mainly forms steep sheets conformable with the pre-existing NNE-trending regional structure on both sides of Qugssuk (Plate 1; Garde, 1989b), although there are also dome-like structures in the northern part of its outcrop area. Both the granitoids were emplaced after the peak of granulite facies metamorphism and associated partial retrogression, after the accretion of tonalitic crust had ceased at least in this part of the Akia terrane.

Granite sheets similar to the Qugssuk granite were also emplaced into an area of *c.* 5 km^2 south-east of Quagssûp taserssua, hosted by granulite facies dioritic gneiss. Like the Qugssuk granite described below, also these granite sheets are medium to fine grained and have granitic textures with dispersed biotite and two feldspars (not mesoperthite as in syn-granulite facies granites). The granite sheets are discordant to gneissic foliation in the dioritic host rock, which is bleached and recrystallised by retrogression along the granite contacts.

Garde *et al.* (1986) and Garde (1989a) presented limited geochemical data indicating that the granitoids are much more evolved than the grey tonalitic gneisses, and approach minimum melt compositions. With support from field evidence and isotopic data these authors and Garde (1990) suggested that the late granitoids were derived by partial melting of source rocks similar to the grey gneiss and Taserssuaq tonalite. Granite geochemistry is discussed on p. 72.

Igánánguit granodiorite dome

The Igánánguit granodiorite dome is a *c.* 20 by 10 km large composite dome located west of Taserssuaq in the south-western part of the Isukasia map area (Plate 1; Figs 47, 48).

The dome has a NNW-trending long axis and is asymmetric: its western flank is much steeper (dipping about 75°) than the eastern flank (dipping 20–40°, Fig. 47). The granodiorite which forms the interior of the dome was emplaced into a host consisting of interlayered and previously folded amphibolite and dioritic to tonalitic orthogneiss. The dome itself is bounded along most of its circumference by a 100–500 m thick layer of amphibolite, which also separates it in the south-east from the Taserssuaq tonalite complex; the position of the north-eastern margin of the dome is only approximately known (S. B. Jensen, personal communication, 1984).

The presence of more or less concentric horizons of amphibolite inside the Igánánguit dome suggests that the granodiorite was in part intruded as separate sheets. In the southern and western parts of the dome, screens of tonalitic gneiss flank internal amphibolite layers, and a 5 km long, bifurcated layer of dioritic gneiss occurs in its central part. The amphibolite and older gneiss horizons and trains of inclusions bring out the internal structure of the dome (Fig 47). It consists of a *c.* 5 km wide northern dome which is separated from a twice as large southern dome by a shallow depression west of Igánánguit. The southern dome itself consists of several smaller domes, which merge southwards into a single south-plunging structure (Fig. 48).

The granitoid rock in the interior of the dome is mostly undeformed but has a distinct schistosity in the immediate vicinity of the older gneiss and amphibolite

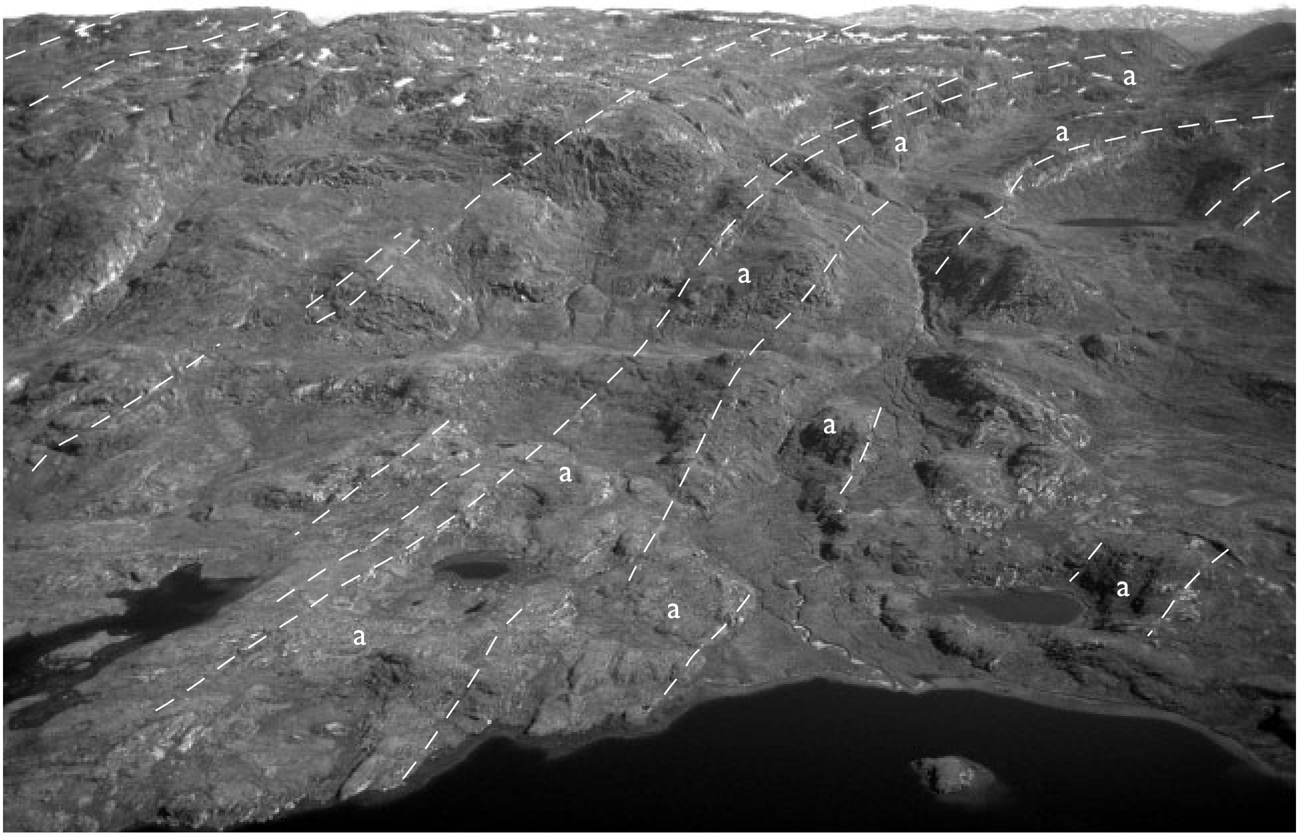

Fig. 48. North-facing slopes in the southern part of the Igánánguit dome. An interior, *c.* 200 m thick amphibolite layer in the central and right parts of the picture (labelled a), which dips away from the observer, outlines the shape of the dome. Other internal boundaries with similar orientations are also marked.

horizons. In spite of this strong foliation its intrusive character into the latter horizons is locally well exposed, and the amphibolites were already strongly deformed before they were intruded by the granite magma. Particularly in the north-western marginal part of the dome the granodiorite commonly contains rafts of amphibolite; some of these rafts are tightly folded with the granodiorite, and the orientations of these (syn-intrusion?) folds suggest shortening perpendicular to the margins of the dome. This is also apparent from Fig. 49 which displays a section perpendicular to the inclined north-western marginal zone of the dome: an amphibolite horizon is intruded by the granite, and the resulting granodiorite-amphibolite agmatite stretched out in directions parallel to the margin.

The Igánánguit granodiorite is a fine- to medium-grained, homogeneous and white to buff or pink-coloured granite to granodiorite (see section on geochemistry for composition), typically with small, evenly dispersed biotite flakes. It locally grades into coarse pegmatitic varieties and is in places cut by white or pinkish pegmatite. Under the microscope the undeformed granite has a more or less equigranular mosaic texture of *c.* 1–3 mm large subhedral grains of quartz, microcline, weakly antiperthitic plagioclase, scattered, *c.* 1 mm large euhedral biotite grains, and occasional small pockets of myrmekitic quartz-feldspar intergrowths. Apatite and zircon are common accessories. There are local signs of minor recrystallisation. The granitic texture with evenly dispersed, small biotite flakes, myrmekite, and absence of mesoperthite show that the Igánánguit granodiorite was emplaced under amphibolite facies conditions and has not been retrogressed from granulite facies. This is taken as independent textural evidence that the Igánánguit granodiorite is younger than the surrounding grey gneiss, most of which has blebby textures indicative of retrogression from granulite facies. Former granulite facies conditions in the area are also indicated by the common presence of orthopyroxene-bearing amphibolite.

As mentioned above there are several structures in the Fiskefjord area that may either be domes formed

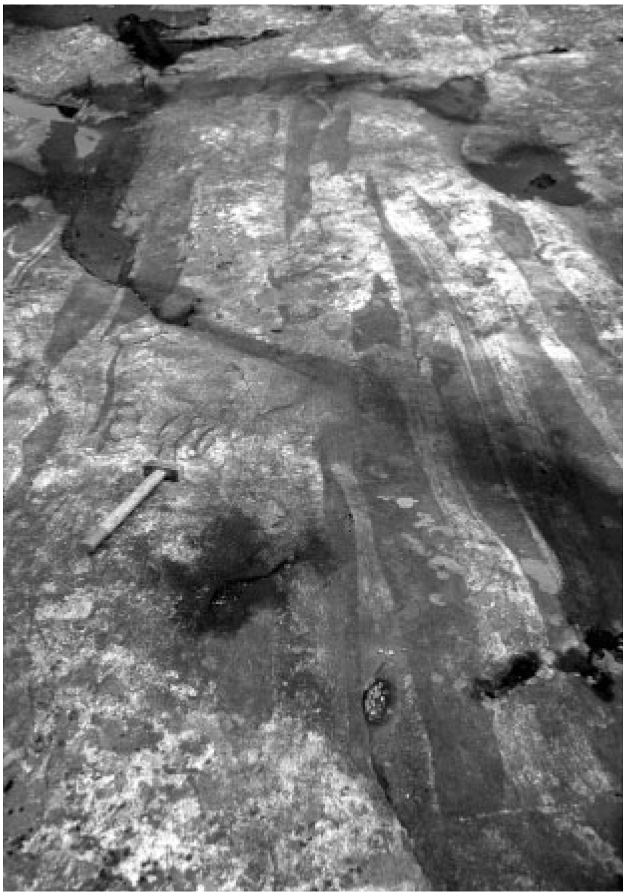

Fig. 49. Deformed marginal zone of the Igánánguit granodiorite with elongate amphibolite rafts. North-western margin of the Igánánguit dome, 9 km west of Taserssuaq.

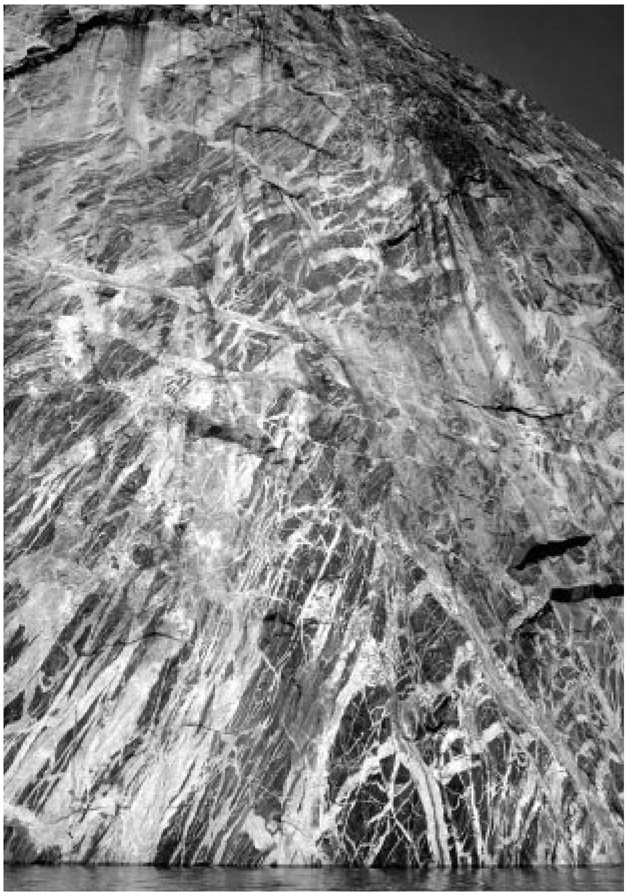

Fig. 50. Amphibolite agmatite with numerous sheets of tonalite, granite and pegmatite (predominantly Qugssuk granite). South-facing cliff opposite the north-eastern end of Bjørneøen.

by injection of magma or diapiric movement, or dome-like structures formed due to superimposed folding. The Igánánguit dome is interpreted as a magmatic or diapiric structure because it consists of a characteristic granitic lithology which is largely undeformed except at its margins, and is younger than the Taserssuaq tonalite complex and multiply deformed grey tonalitic gneiss which it intrudes.

The Igánánguit granodiorite is about 3000 Ma old. Field relationships and metamorphic textures presented here show that it is younger than the surrounding grey gneiss and Taserssuaq tonalite complex, but Rb-Sr and Pb-Pb isotopic investigations have given very imprecise results. Field observations revealed that the Igánánguit granodiorite commonly contains millimetres to centimetres thick biotite-rich wisps which most likely represent restite material from its source, and the possibility must be considered that the granodiorite is not isotopically homogeneous. Regression of 12 samples analysed for Rb-Sr gave an apparent age of 2935 ± 240 Ma (initial $^{87}Sr/^{86}Sr$ = 0.7021 ± 0.004, MSWD = 9.6). The best fit for six of these samples, collected within an area of 1 km^2, was 3013 ± 190 Ma (initial $^{87}Sr/^{86}Sr$ = 0.7013 ± 0.0008, MSWD = 1.98) (Garde *et al.*, 1986). Pb isotopic analysis (P. N. Taylor, personal communication, 1986) resulted in a well-fitted isochron of 3092 ± 48 Ma with a model μ1 value of 7.56 for 11 samples (μ1 values around 7.5 are normal for Middle Archaean orthogneisses in West Greenland). However, this result is heavily dependent on a single very radiogenic sample (GGU 278880). Omitting this sample the age is 3023^{+195}_{-227} Ma, model μ1 = 7.47. The Pb-Pb isotopic data are further discussed on p. 79.

Qugssuk granite

The Qugssuk granite (Garde, 1984; Garde *et al.*, 1986) straddles the boundary between grey gneiss retrogressed from granulite facies and the Taserssuaq tonalite

Fig. 51. Qugssuk granite (left) cutting strongly deformed, retrogressed grey gneiss with small elongate amphibolite enclaves. See discussion in the main text. North coast of Qugssuk.

pluton north and east of Qugssuk. It is a medium-grained, white to pale pinkish leucocratic biotite granite which may grade into coarse pegmatitic varieties.

A whole-rock Rb-Sr isochron age of 2969 ± 32 Ma, initial $^{87}Sr/^{86}Sr$ = 0.7020 ± 0.0003, MSWD = 1.09, was obtained by Garde *et al.* (1986). A younger but less precise Rb-Sr age was obtained from eight samples collected from granite sheets on the peninsula east of Qugssuk. These data points plot along a best-fit line with an age of 2842 ± 85 Ma, with a relatively high initial $^{87}Sr/^{86}Sr$ ratio of 0.7040 ± 0.0009, and MSWD = 6.32 (Fig. 38c; Table 2). Pb isotopic data discussed by Garde (1989a, 1990) and on p. 79 suggest open-system behaviour of lead.

Unlike the Igánánguit granodiorite, the Qugssuk granite has been emplaced mainly as anastomosing sheets up to a few metres or tens of metres thick, generally with steep or subvertical orientations roughly parallel to the orientations of the host rocks (Fig. 50). The granite also forms much thicker (up to *c.* 1 km) sheets that have been intruded into the marginal zone of the Taserssuaq tonalite complex east of Qugssuk, and into a major amphibolite unit north of Qugssuk. Particularly in the latter area there are large, homogeneous outcrops devoid of older enclaves, and here a couple of small dome-like structures occur.

Fig. 51 shows an example of the intrusive relationship between a sheet of Qugssuk granite and retrogressed grey gneiss at the north coast of Qugssuk. The interpretation by Garde (1990, p. 670) that the granite cuts pre-existing retrogressive textures in the grey gneiss may be true; alternatively the granite itself was the cause of retrogression, which was brought about by hydrous fluids liberated during solidification of the granite (see p. 77 ff.).

Terrane assembly

The Archaean crust in the Godthåbsfjord region consists of several terranes with different ages, lithologies and tectonic and metamorphic histories (Friend *et al.*, 1988a); the Akia and Akulleq terranes and their common tectonic boundary (Fig. 2) were first described by Friend *et al.* (1988b). Their assembly is the latest Archaean event known to have affected the Akia terrane. The Ivinnguit fault that forms the terrane boundary is a NE-trending, subvertical to moderately WNW-dipping, 10–15 m thick mylonite zone, which is best exposed on the coasts of south-eastern Bjørneøen and Sadelø in outer Godthåbsfjord (McGregor *et al.*, 1991). The terrane boundary straddles the south-eastern corner of the Fiskefjord map area (Fig. 2).

The timing of terrane assembly is poorly constrained but occurred in the interval between *c.* 2800 Ma (the age of Ikkatoq gneisses in the Akulleq terrane which are cut by the Ivinnguit fault) and *c.* 2720 Ma (the ages of granite sheets that cut several terrane boundaries, McGregor *et al.*, 1991; Friend *et al.*, 1996). Thus, a SHRIMP zircon U-Pb age of 2712 ± 9 Ma was reported by Friend *et al.* (1996) from a Qârusuk dyke at the type locality on Bjørneøen close to the south-eastern margin of the Akia terrane. However, no such dykes have so far been observed in the Fiskefjord area, and it is uncertain how far the effects of the terrane assembly can be traced north-west of the terrane boundary itself.

McGregor *et al.* (1991) and McGregor (1993) described a NNE–SSW zone of strong ductile deformation continuous with the Qugssuk–Ulamertoq high strain zone along western Godthåbsfjord, which according to these authors postdates the juxtaposition of the Akia and Akulleq terranes. However, as discussed above, at least the Qugssuk–Ulamertoq high strain zone itself was established prior to the assembly of the two terranes.