having had a greater distance available to rise in the crust, would generally be more effectively separated from restite minerals (and hence have the more felsic composition of the two groups).

Except for Cr contents which are always very low (1–20 ppm) all granite groups in the Fiskefjord area would chemically belong to group CA_2 of Sylvester (1994), with the mesoperthite granite providing the best fit. It seems reasonable from granite field relationships and *P-T* estimates of the granulite facies metamorphism in the southern part of the Akia terrane (Reed, 1980; Pillar, 1985; Riciputi *et al.*, 1990) that the granites could have been melted from sources at pressures up to *c.* 10 kbar. In addition, the fact that the granites have more fractionated REE patterns than the grey gneiss supports the idea that hornblende ± garnet were important restite phases, assuming that the granitic melts were indeed separated from a source of grey gneiss.

Granulite facies metamorphism, retrogression and element mobility in grey gneiss

Granulite facies metamorphism

Granulite facies metamorphism extends over the south-western part of the Akia terrane, including Nordlandet and a large part of the Fiskefjord area, whereas the easternmost part of the Fiskefjord area (and of the Akia terrane) consists of upper amphibolite facies rocks that have not experienced granulite facies metamorphism. Large tracts between these two areas are variably retrogressed (Fig. 76). Whole-rock Pb-Pb ages of 3000 ± 70 Ma (Taylor *et al.*, 1980) and 3112 ± 40 Ma (Garde, 1989a) obtained from orthogneiss in Nordlandet and the eastern part of the Fiskefjord area have been interpreted to date the granulite facies metamorphism; a 2999 ± 4 Ma SHRIMP age of metamorphic zircon overgrowth in a garnet-sillimanite-bearing metasediment from Nordlandet just south of the Fiskefjord area (Friend & Nutman, 1994) firmly establishes that the peak of granulite facies metamorphism occurred at *c.* 3000 Ma, i.e., it culminated during or just after emplacement of the main phase of grey tonalitic-trondhjemitic gneiss. SHRIMP zircon data from Nordlandet dioritic gneiss reported in Table 1 and Fig. 25 suggest that this unit also experienced an earlier thermal event at *c.* 3180 Ma. The *c.* 3000 Ma granulite facies metamorphism outlasted two phases of isoclinal folding and a phase of upright, more open folding (the Pâkitsoq phase, Berthelsen, 1960) in the western and central parts of the Fiskefjord area. Granulite facies metamorphism appears to have overlapped with doming, and was succeeded by localised ductile deformation and emplacement of small granodiorite and granite plutons and granite sheets in the northern and eastern parts of the area.

Physical conditions of metamorphism

In Nordlandet the peak of metamorphism occurred under *P-T* conditions of *c.* 850°C, 8 kbar (Reed, 1980; Riciputi *et al.*, 1990), and at Langø, a small island west of Tovqussap nunâ, *c.* 825°C and 8.3 kbar were reported by Dymek (1984). Titanium-rich hornblende and biotite appear to have been stable throughout the granulite facies event in suitable rock types (Garde, 1990 and this paper), indicating that complete dehydration did not take place. According to Pillar (1985) and Riciputi *et al.* (1990) granulite facies metamorphism took place without free fluids, and fluid inclusion data from grey gneiss in the Fiskefjord area seem to support this (Garde, 1990). However, some aspects of the granulite facies rocks may be ascribed to fluid activity during the granulite facies event, for instance hornblende-bearing mafic pegmatites on Nordlandet in which 'high-grade' hornblende partially replaces orthopyroxene (McGregor *et al.*, 1986; McGregor, 1993); also the geochemistry of granulite facies biotite (p. 63) may suggest the former presence of metamorphic fluids (see p. 78).

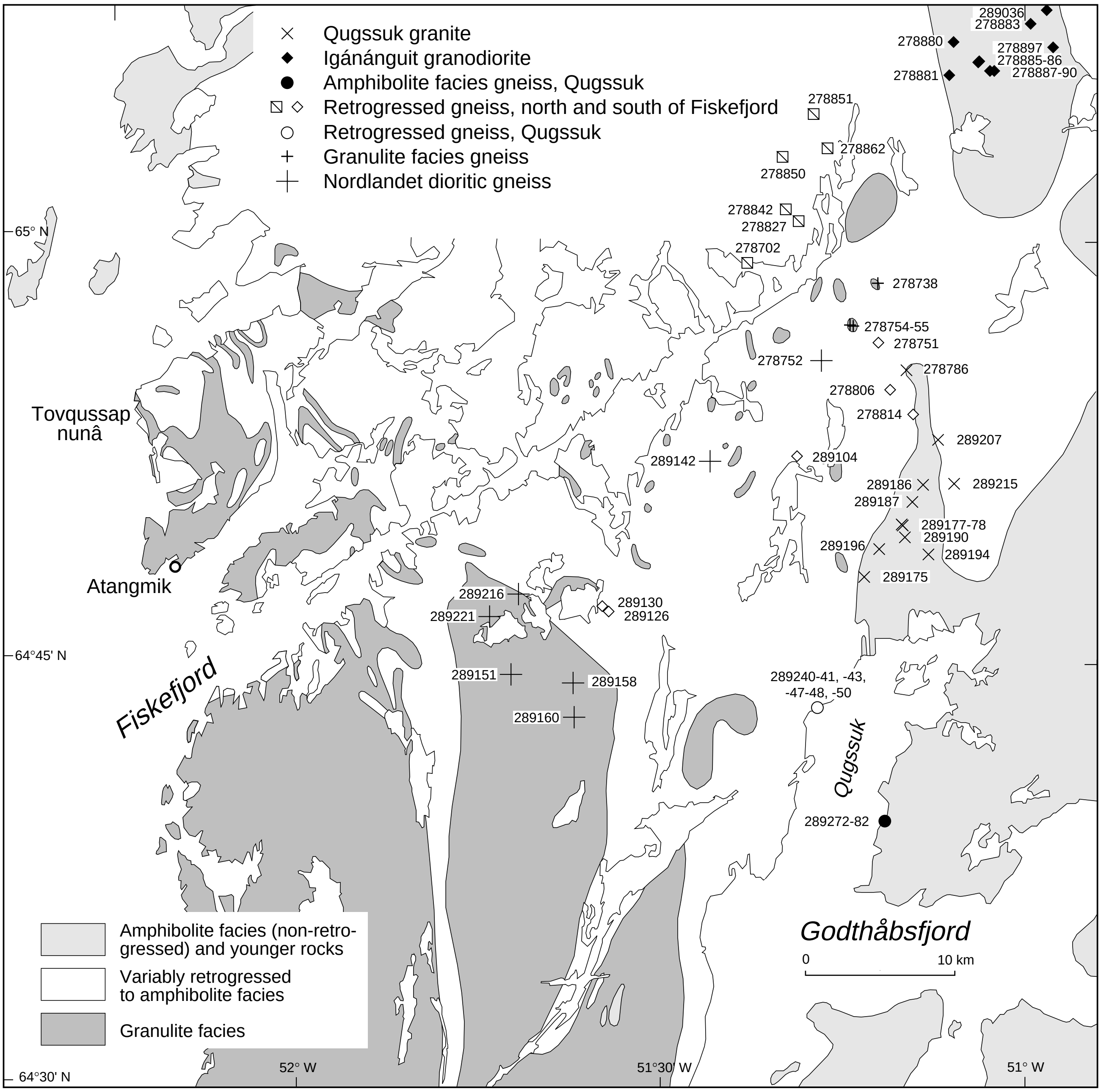

Fig. 76. Distributions of granulite facies, amphibolite facies (not retrogressed from granulite facies), and variably retrogressed grey gneiss in the Fiskefjord area modified from Garde, 1990, fig. 2, and locations of samples used for Pb-Pb and Rb-Sr isotope geochemical study. The amphibolite facies areas comprise both grey gneiss east of Qugssuk, the Finnefjeld gneiss complex in the north-west, the Igánánguit granodiorite in the north-east, and the Taserssuaq tonalite complex and Qugssuk granite north and east of Qugssuk. Samples used for Pb-Pb isotope geochemistry (Fig. 78) are: Qugssuk granite, Igánánguit granodiorite, amphibolite facies and retrogressed gneiss at Qugssuk, and Nordlandet dioritic gneiss (including GGU 289142 and 278752 north-east of the main outcrop). Samples used for Rb-Sr isotope geochemistry and shown in Fig. 79 are: amphibolite facies and retrogressed gneiss from all localities, and Qugssuk granite. Rb-Sr geochronology of the Igánánguit granodiorite is discussed on p. 47.

Phases of metamorphism

Dymek (1978, 1984), who studied metamorphism of supracrustal rocks from several parts of the Godthåbsfjord region (within the Fiskefjord area including rocks from Langø – Tovqussap nunâ and Qugssuk), described a regional upper amphibolite to hornblende granulite facies metamorphic event *M1* and a regional retrogressive event *M2*, the latter within the stability field of kyanite. Although his study was carried out before it was realised that the Godthåbsfjord region consists of several terranes with different magmatic, tectonic and metamorphic histories, his observations regarding the *M2* metamorphic phase are pertinent to the Fiskefjord area. Dymek (1978, 1984) noted that the *M2* event was very variably developed within a given outcrop or even within a single thin section. When Dymek (1978) first described the *M2* event, no source for the thermal input or water necessary for hydration was proposed. In a subsequent paper Dymek (1984) suggested a possible correlation of the *M2* event with shear heating and hydration along predominantly subvertical high-grade shear zones of local and regional extent.

Cause of granulite facies metamorphism

Garde (1990) discussed several possible mechanisms of the *c.* 3000 Ma granulite facies metamorphism in the Fiskefjord area (emplacement of originally dry magma; dehydration by CO_2 streaming; thermal metamorphism) and in agreement with other workers cited above concluded that the *c.* 3000 Ma granulite facies metamorphism was probably caused by heat accumulated during continuous injection of tonalitic magma into the growing continental crust. This mechanism of thermal metamorphism by over-accretion had previously been described in detail by Wells (1979, 1980) and applied to *c.* 2800 Ma granulite facies metamorphism in the Buksefjorden area south of Godthåbsfjord (see also Garde, 1990, p. 679).

Retrogression and element mobility

Garde (1989a, 1990, this paper) described progressive changes on outcrop and microscopic scale during prograde and retrograde metamorphism of grey gneiss and presented evidence of element mobility during metamorphism. He showed that in the northern and central parts of the Fiskefjord area most of the grey gneiss (and to a lesser extent mafic and ultramafic supracrustal rocks) has been partially or completely retrogressed from granulite facies, mainly under static conditions, has disequilibrium mineral assemblages and textures, and contains amphibole and biotite which formed over a wide range of amphibolite facies *P-T* conditions – in places two secondary generations of biotite. Only the eastern and northernmost parts of the Fiskefjord area escaped granulite facies metamorphism. Here the orthogneiss contains upper amphibolite facies metamorphic parageneses with equilibrium mineral textures. The original prograde amphibolite to granulite facies boundary in grey gneiss is not preserved in the Fiskefjord area, having been overprinted by retrogression. In West Greenland, outcrops of prograde amphibolite to granulite facies transitions in grey gneiss (resembling those in e.g. the Kabbaldurga quarry, South India, Pichamuthu, 1960) have so far only been described from the southern part of the Fiskenæsset region south of Nuuk, within the Tasiusarsuaq terrane (McGregor & Friend, 1992).

Timing and significance of retrogression, and mechanisms of element transport

Field observations, conventional and isotope geochemistry, mineral chemistry, and a *c.* 3000 Ma U-Pb zircon age obtained from a post-kinematic diorite plug emplaced during or after retrogression suggested to Garde (1990, 1991) that much of the retrogression took place very soon after the culmination of granulite facies metamorphism. Garde (1990) proposed a mechanism of retrogression whereby continuous dehydration and partial melting under granulite facies *P-T* conditions at depth were accompanied by penecontemporaneous retrogression slightly higher in the crust. Liberation of aqueous fluids from the hydrous anatectic melts, as they moved into the upper levels of granulite facies orthogneiss and solidified, would lead to partial or complete retrogression of previously dehydrated rocks, as part of the same thermal event. Figure 77 schematically shows several stages of this development, whereby dehydration and development of a diffuse anatectic network were followed by local mobilisation of leucocratic melts and recrystallisation with secondary amphibole and biotite. Leucocratic veins only centimetres thick may not have travelled far from their source, but a similar process may have operated at a scale of metres to a few tens of metres, as evidenced by retrogression of dioritic gneiss immediately adjacent to granite sheets south-east of Quagssûp taserssua. In the case of the Qugssuk granite, which was emplaced not long after granulite facies metamorphism and is likely to have been derived by coalescing anatectic melts formed during the 3000 Ma granulite facies metamorphism (Rb-Sr whole-rock ages 2969 ± 32 Ma and 2842 ± 85 Ma, Tables 1, 2; Fig. 38), no such halo of ret-

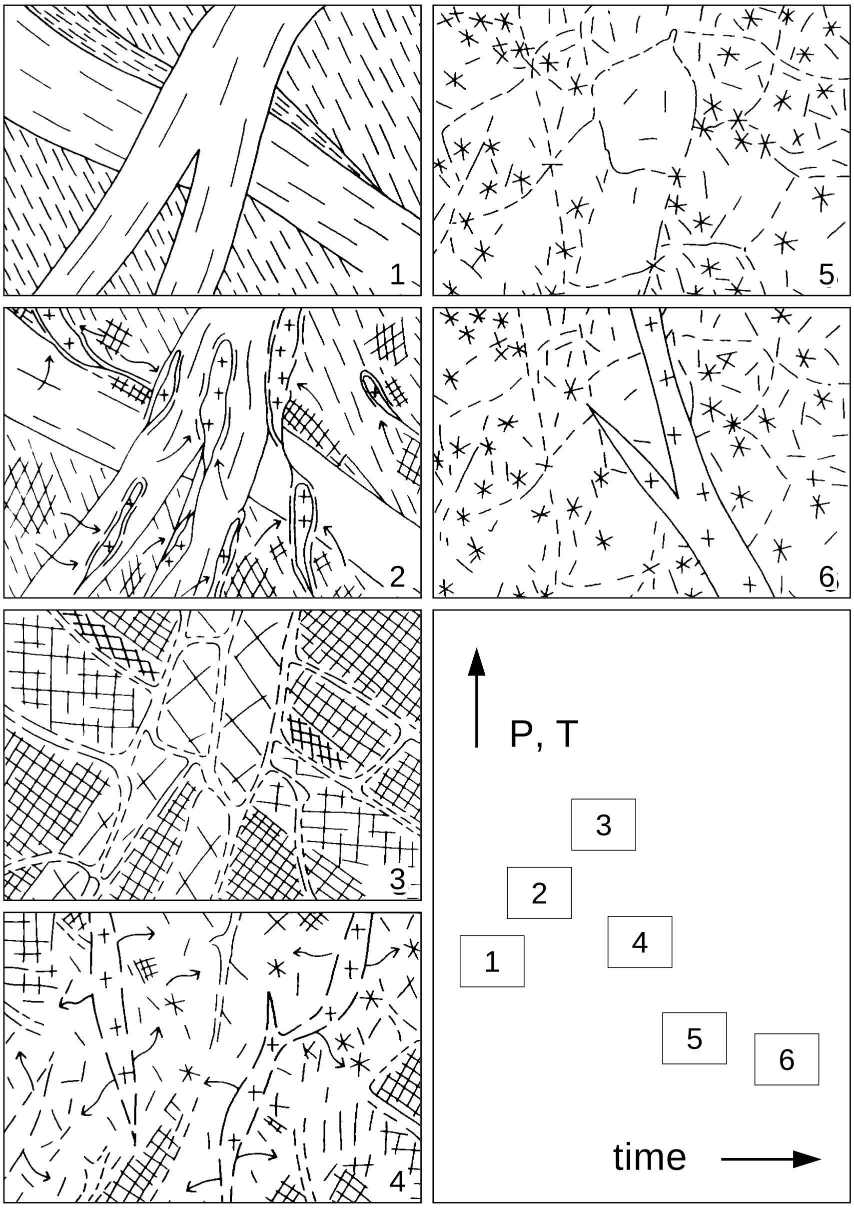

Fig. 77. Schematic stages of grey gneiss development in the Fiskefjord area during granulite facies metamorphism and 'high-grade' retrogression (modified from Garde, 1990, fig. 16). (**1**) Polyphase amphibolite facies gneiss, as shown in Fig. 31. (**2**) Early stage of dehydration and growth of orthopyroxene in cross-hatched areas, and beginning of partial melting (++). (**3**) Granulite facies stage with diffuse quartz-plagioclase anatectic network. Compare Figs 26 (undeformed) and 27 (deformed). (**4**) Early stage of retrogression; former orthopyroxene grains become visible as secondary amphibole-biotite blebs (stars) (compare Figs 32 and 45). (**5**) Fully retrogressed and recrystallised gneiss with blebby texture, compare Fig. 33. (**6**) Retrogressed gneiss intruded by syn- or post-retrogression granite (compare Fig. 51). For reasons unknown, the orthopyroxene crystals (viz. blebby texture) may preferentially be located in the quartzo-feldspathic veins (Fig. 32), in the surrounding rock (Fig. 27), or in both positions (Fig. 33).

rogression can be identified. The granite sheets are bounded to the west and north by retrogressed grey gneiss which, except for islands of granulite facies gneiss, extend many kilometres to the west, north and north-east. Garde (1990) argued that the Qugssuk granite cuts, and is hence younger than, the blebby textures in grey gneiss north of Qugssuk; alternatively V. R. McGregor suggested (personal communication, 1995) that the retrogression in the adjacent tonalitic-trondhjemitic gneiss could have been caused by fluids emanating from the granite during its solidification. This is further discussed below.

It was demonstrated in previous sections that both granulite facies metamorphism and retrogression appear to have been accompanied by very significant migration of LIL elements in the orthogneisses. The migrating elements may have been transported both in anatectic quartzo-feldspathic melts as suggested by field observations and bulk geochemistry, in concentrated fluids (perhaps coexisting with the melts) as suggested by low Rb contents of granulite facies biotite, and in aqueous fluids released from the crystallising quartzo-feldspathic melts. It was also shown that retrogression took place under a large range of upper to lower amphibolite facies *P-T* conditions, as evidenced by disequilibrium mineral textures and the compositions of retrograde amphibole and biotite (Garde, 1990).

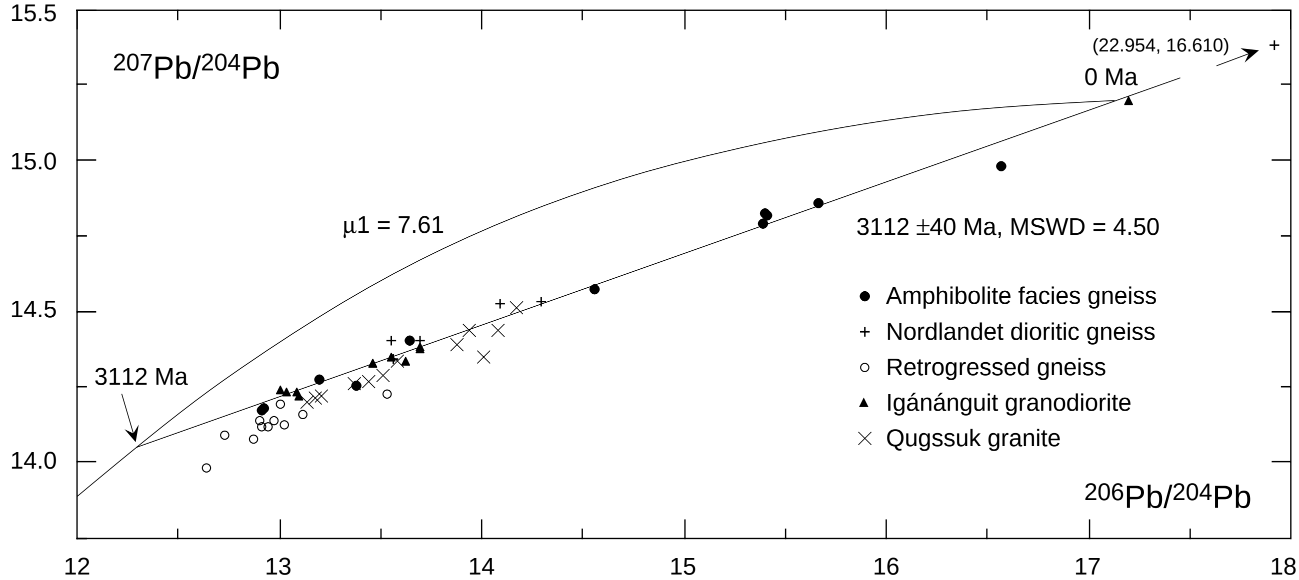

Fig. 78. Pb-Pb isochron diagram of grey gneiss and granitoid rocks from the Fiskefjord area (modified from Garde, 1990, fig. 13). The 3112 Ma age is calculated from samples of amphibolite facies grey gneiss at Qugssuk, Igánánguit granodiorite and Nordlandet dioritic gneiss (see also p. 28). The retrogressed grey gneiss and the Qugssuk granite both plot below the 3114 Ma line and have been contaminated with less radiogenic lead. Sample locations are shown in Fig. 76 (all samples of retrogressed gneiss were collected on the north-west coast of Qugssuk). Analyst: P. N. Taylor.

In the Rb-depleted granulite facies samples from which minerals were analysed, titanium-rich high-grade (granulite facies) biotite with low Rb contents occurs in equilibrium with hornblende and hypersthene (Tables 3, 6). Morphologically (and presumably compositionally) similar high-grade biotite is common in many other granulite facies samples. In theory, this refractory biotite, like phlogopite, should retain Rb in preference to a granitic melt, and the observed Rb distribution is therefore not well explained by granitic melt extraction. Hansen & Newton (1995) described a similar pattern of inverse correlation between Rb and TiO_2 in biotite from an area of prograde amphibolite to granulite facies transition in southern Karnataka, India. They interpreted the low Rb content of granulite facies biotite as due to Rb extraction by a pervasive fluid with high partitioning of Rb relative to biotite under granulite facies *P-T* conditions – perhaps a concentrated, immiscible chloride-carbonate brine coexisting with a quartzofeldspathic melt.

Pb and Rb-Sr isotope data: further evidence of mechanisms and timing of retrogression

Pb and Rb-Sr isotopic data from grey gneiss, Igánánguit granodiorite and Qugssuk granite in the eastern part of the Fiskefjord area were reported by Garde (1989a, 1990). Isotopic ages were discussed in the section on magmatic accretion, and additional Rb-Sr data are presented here. Regarding the U-Pb system, Garde (1990) presented Pb-Pb whole rock data from grey gneiss and granite (P. N. Taylor, personal communication, 1990) showing that the lead isotopic compositions of retrogressed gneiss and Qugssuk granite are less radiogenic than those of the other groups (Fig. 78; locations of samples Fig. 76). This pattern is evidence of open-system behaviour of lead and also suggests contamination with (?Early Archaean) unradiogenic lead during retrogression. The data also show that the lead source of the Qugssuk granite was not isotopically homogenised.

As regards the Rb-Sr system, 10 samples of amphibolite facies gneiss (not retrogressed from granulite facies) define an errorchron of 2954 ± 120 Ma (initial $^{87}Sr/^{86}Sr$ = 0.7014 ± 0.0004, MSWD = 6.85) (Fig. 79a and Garde, 1989a; sample locations Fig. 76). Retrogressed samples from the north-western coast of Qugssuk plot along this line but very near its origin (Fig. 79b). Other granulite facies and retrogressed samples of grey gneiss from the north-eastern part of the Fiskefjord area (Table 7) plot a little above the 2954 Ma reference line but parallel to it, despite some scatter in the data points (Fig. 79c). A regression of these points gives an apparent age of 3137 ± 172 Ma, initial $^{87}Sr/^{86}Sr$ = 0.7017 ± 0.0001, and

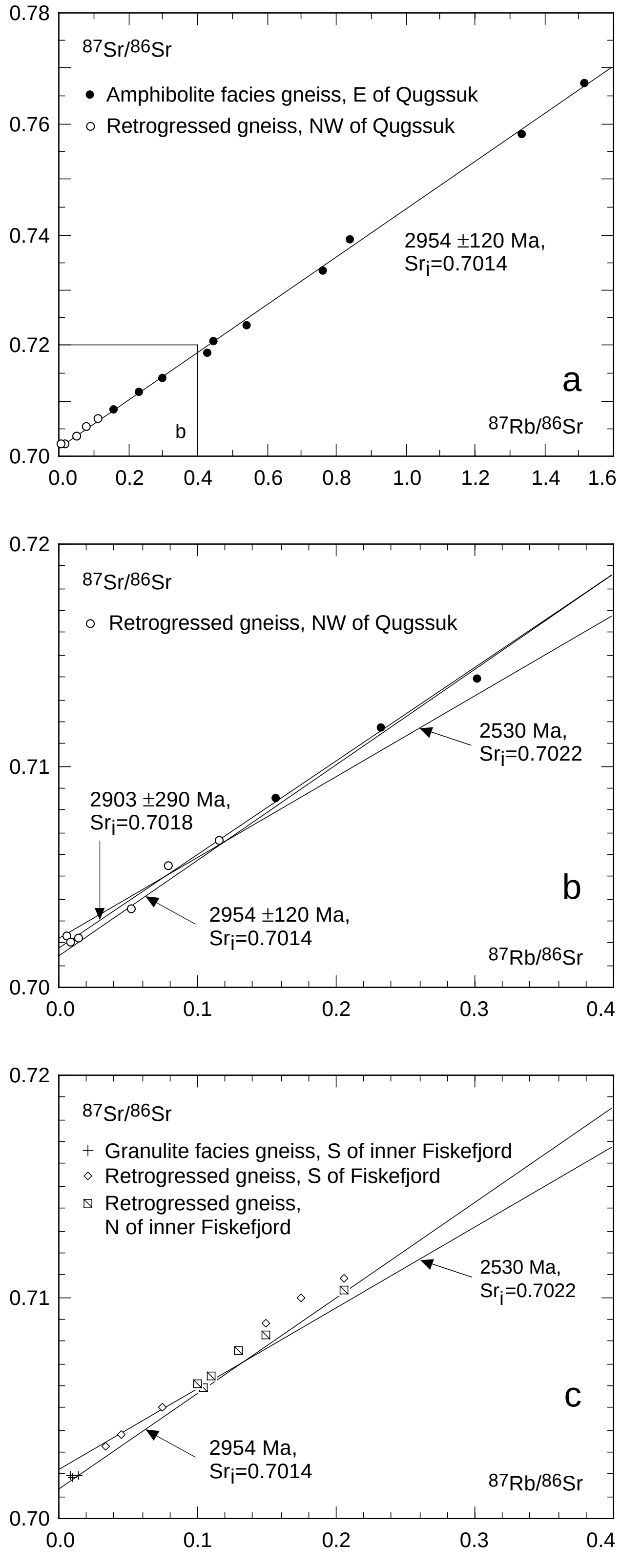

Fig. 79. Rb-Sr isochron diagrams for grey gneiss around Qugssuk and inner Fiskefjord. Sample locations are shown on Fig. 76. (**a**) Amphibolite facies gneiss 289272–289282 from the eastern side of Qugssuk with a regression line (2954 ± 120 Ma (2σ), Sr_i = 0.7014 ± 0.0003, MSWD = 6.85 for these points. Retrogressed gneiss from the western side of Qugssuk (open circles) plots on the same line but near its origin (error magnification is used in the calculation of all 2σ values; modified from Garde, 1990, fig. 12). (**b**) Enlarged portion of (a) with an errorchron of 2903 ± 290 Ma (2σ), Sr_i = 0.7018 ± 0.0004 for the retrogressed gneiss samples NW of Qugssuk, and two reference lines of 2954 Ma and 2530 Ma (estimated Sr_i = 0.7022) for comparison (see the main text). (**c**) Rb-Sr isochron diagram for granulite facies and retrogressed grey gneiss located around inner Fiskefjord. The diagram shows data points from three groups of samples (Table 7): granulite facies and retrogressed grey gneiss from the area south of Fiskefjord, and a group of retrogressed, very leucocratic gneiss occurring north of inner Fiskefjord; several samples lie within the Qugssuk–Ulamertoq zone. The data points from both granulite facies and retrogressed gneiss clearly show that their Rb-Sr isotope systems were closed quickly after 3000 Ma ago. In spite of some disturbance due to granulite facies metamorphism and retrogression (and perhaps minor initial isotopic inhomogeneity) their initial $^{87}Sr/^{86}Sr$ ratio remains low, and their apparent age is not lowered; regression of the data points shown in Table 7 and Fig. 79c gives an apparent age of 3137 ± 192 Ma, initial $^{87}Sr/^{86}Sr$ = 0.7016 ± 0.0001, and MSWD = 2.20.

MSWD = 2.20. If significant migration of Rb and Sr took place during retrogression (as is strongly suggested by Rb and Sr concentration data, petrography and mineral compositions reported in previous sections), this must have happened soon after closure of the Rb-Sr system in its igneous precursor.

McGregor (1993) suggested that much of the retrogression in the eastern part of the Fiskefjord area was related to major Late Archaean (*c.* 2530 Ma?) deformation along the north-western coast of Qugssuk and the Qugssuk–Ulamertoq zone (see p. 82), with potential disturbance of the Rb-Sr system which would have affected most or all of retrogressed gneiss samples referred to above.

Field observations by the author in the Qugssuk–Ulamertoq zone itself only suggest limited Late Archaean deformation. Besides, disturbance of an isotopically homogeneous Rb-Sr system significantly later than 3000 Ma would have resulted in a decrease of the apparent age of the retrogressed gneiss, and would also have increased the apparent initial $^{87}Rb/^{86}Sr$ ratio in the cases of later isotopic homogenisation, or later addition of Rb or crustal Sr. Such changes would be difficult to detect in the retrogressed gneiss on the north-west coast of Qugssuk due to their low Rb/Sr ratios, but they would be apparent in the other groups south and north of inner Fiskefjord. For instance, if isotopic homogenisation of these retrogressed gneisses took place in a closed system at 2530 Ma, an initial $^{87}Sr/^{86}Sr$ ratio of 0.7014 at 2954 Ma would have increased to *c.* 0.7022 at 2530 Ma (using an average $^{87}Rb/^{86}Sr = 0.123$). On Fig. 79c the data points of granulite facies and (mainly) retrogressed gneiss from the two latter areas are shown together with two reference lines of 2954 and 2530 Ma. The initial $^{87}Sr/^{86}Sr$ ratio of the retrogressed gneiss (calculated value 0.7017 ± 0.0001) actually does appear to be slightly higher than that of amphibolite facies gneiss (initial ratio 0.7014 ± 0.0004), but the apparent difference is very small and within analytical error. In addition, the age is not reset; an age older than 2954 Ma is actually indicated, as noted above.

Significance of blebby texture

The two alternative interpretations of the field relationship between Qugssuk granite and retrogressed gneiss presented above prompt a discussion of the significance of blebby texture, commonly observed in retrogressed gneiss whatever the cause of its retrogression. V. R. McGregor (personal communication, 1995) pointed out that a major part of the textural modification observable in outcrops of retrogressed grey gneiss is likely to have formed during the preceding granulite facies metamorphic event, whereby iron and magnesium derived from the evenly distributed hornblende or biotite characteristic of not previously retrogressed amphibolite facies gneiss are concentrated in fewer and larger orthopyroxene crystals or crystal aggregates (or garnet in rocks of appropriate composition). This (granulite facies) textural coarsening of mafic components only becomes apparent in the field in the form of blebby texture if and when the rocks are rehydrated, and felsic and mafic minerals again become easily recognisable by colour. It may therefore be difficult to establish the relative timing of granite emplacement and retrogression of an adjacent body of older rocks on textural grounds alone (Fig. 51). However, in the case of the Qugssuk granite, the age of granite emplacement would provide a minimum age of retrogression, irrespective of whether retrogression was caused by the granite or had already taken place.

Causes of retrogression

Retrogression of high-grade terrains is conventionally explained by later geological events which are unrelated to granulite facies metamorphism, for instance along major thrusts or shear zones, whereby water is introduced at a higher crustal level from underthrust hydrated rocks. This was, e.g. suggested for the northern boundary of the Tasiusarsuaq terrane in the Godthåbsfjord region (Friend *et al.*, 1988b).

Was retrogression in the Fiskefjord area mainly related to granulite facies metamorphism as suggested by Garde (1990) and outlined above, or was it mainly caused by various younger unrelated events? In the Fiskefjord area later retrogression of country rock gneiss might be surmised to happen during emplacement of post-granulite facies Archaean plutons such as the Finnefjeld gneiss complex, adjacent to faults at the time of juxtaposition of the Akia and Akulleq terranes, during Late Archaean ductile 'straight belt' deformation, and related to Early Proterozoic dykes and faults (McGregor, 1993; V. R. McGregor, personal communication, 1995; McGregor *et al.*, 1991). A fully satisfactory answer to these questions is difficult to obtain, although the very variable nature of retrogression as displayed by mineral assemblages, mineral chemistry and microscopic textures may point to high-grade and low-grade retrogression operating successively and with variable intensity in the same areas.

Table 7. Rb-Sr whole rock data for grey gneiss around inner Fiskefjord

	Rb ppm	Sr ppm	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
Granulite facies gneiss, south of inner Fiskefjord				
278738	1.7	568	0.009	0.7019
278754	3.8	540	0.014	0.7020
278755	2.0	543	0.010	0.7018
Retrogressed gneiss, south of Fiskefjord				
278751	29	456	0.175	0.7100
278806	28	381	0.206	0.7109
278814	32	581	0.149	0.7089
289104	14	923	0.046	0.7038
289126	14	1050	0.034	0.7033
289130	24	922	0.075	0.7050
Retrogressed gneiss, north of inner Fiskefjord				
278702	52	1110	0.130	0.7076
278827	34	937	0.105	0.7059
278842	45	883	0.150	0.7083
278850	24	681	0.100	0.7061
278851	43	590	0.205	0.7103
278862	32	843	0.110	0.7065
283379	147	132	3.259	0.8393

See discussion in the main text and Fig. 79. Analytical methods as described by Garde *et al.* (1986); the precision of Rb/Sr measurements is within *c.* 1% (2σ) for samples with more than *c.* 5 ppm Rb, and of $^{87}Sr/^{86}Sr$ measurements better than 0.0002 (2σ).

The south-eastern boundary zone of the Finnefjeld gneiss complex adjacent to grey and purple gneiss was studied by Marker & Garde (1988); contact relationships are complicated, with several different intrusive phases; in places, granulite facies gneiss occurs quite close to the margin of the complex. The northern and interior boundary zones of the complex are not well known, but earlier reconnaissance mapping indicates that granulite facies gneiss occurs at or close to the margins of the complex in these areas (Allaart, 1982). In the view of the author, known field relationships do not suggest that the emplacement of the Finnefjeld gneiss complex caused widespread retrogression of the surrounding grey gneiss – at least not at the exposed level.

Retrogression along Late Archaean ‘straight belts’ of ductile deformation was observed by McGregor *et al.* (1991) and McGregor (1993) in a number of places in the Godthåbsfjord region, a possibility also mentioned by Dymek (1984). One of these belts, supposed to be *c.* 2530 Ma old (McGregor, 1993), occurs along the west coast of outer Godthåbsfjord and continues into the Qugssuk–Ulamertoq zone. Along with other ‘straight belts’ in the Qugssuk area McGregor envisaged it to be a major cause of lateral fluid movement at a scale of kilometres and concomitant retrogression of grey gneiss in the eastern part of the Fiskefjord area. Contrary to McGregor’s interpretation, it was shown above that disturbance of Rb-Sr isotopic systems of this retrogressed grey gneiss must have happened close to 3000 Ma ago. It was also shown above that the pervasive N–S structural grain of the Qugssuk–Ulamertoq zone, including the transposition of earlier folds into upright isoclines, was acquired while granulite facies conditions still prevailed. Far from precluding localised Late Archaean ductile deformation and fluid movement along the Qugssuk–Ulamertoq zone, this observation merely shows that late reactivation was not regionally important in this area. Besides, notwithstanding that pro- and retrograde granulite facies transitions are to some extent lithologically controlled (see e.g. discussion in Garde, 1990 p. 670), substantial parts of the more leucocratic (tonalitic-trondhjemitic) tracts in eastern Nordlandet south of the Fiskefjord area would supposedly also have been affected by retrogression, if Late Archaean lateral fluid infiltration at a scale of many kilometres had taken place along this zone. This is not the case.

Effects of Proterozoic retrogression can be observed in the field along mafic dykes and faults, but the retrogression is restricted to narrow subvertical zones a few metres wide, or at the most a few hundred metres wide along the Fiskefjord fault (a narrow fjord-parallel zone of retrogression occurs in granulite facies gneiss where the fault comes closest to the shores in outer Fiskefjord). However, the Sr isotopic composition of retrogressed gneiss shows that Proterozoic retrogression did not have regional significance.

Conclusions

The widespread retrogression in the central and eastern parts of the Fiskefjord area had several causes. Some Late Archaean and Proterozoic retrogression unrelated to granulite facies metamorphism is readily visible at small scale along late shear zones, faults, and at the margins of Proterozoic dykes. However, the author contends that much of the retrogression is best explained as related to *c.* 3000 Ma granulite facies thermal metamorphism. In spite of difficulty in assessing the precise compositions of magmatic precursors to grey gneiss there is strong geochemical evidence that both granulite facies metamorphism and retrogression were accompanied by mobility of LIL elements. The presence of granulite facies quartzo-feldspathic veins and syn- and post-granulite facies I-type granites suggest that dehydration melting of grey gneiss took place. Fluid activity at granulite facies conditions is indicated from Rb geochemistry of biotite. Petrography and mineral chemistry suggest that retrogression took place under a range of upper to lower amphibolite facies P-T conditions, and that there were perhaps two or more episodes of retrogression in some areas. Field relationships between the Qugssuk granite and adjacent grey gneiss may be equivocal in terms of the age of retrogression, but Rb-Sr isotope geochemistry of grey gneiss in this part of the Fiskefjord area independently indicates that a large part of the retrogression took place not later than *c.* 2950 Ma ago. There is field evidence that also Late Archaean and Proterozoic retrogression not related to granulite facies metamorphism took place, but the author has not found evidence of *widespread, Late* Archaean deformation along the Qugssuk–Ulamertoq zone or other zones of high strain in the Fiskefjord area, or of accompanying fluid activity and retrogression with regional importance.

Post-kinematic diorites

Field and petrographic observations

About twenty individual bodies of post-kinematic diorite range in outcrop size from a few square metres to *c.* 1 km^2 (Garde, 1991, fig. 1), but more may be present; they are not easily recognised in areas of granulite facies tonalitic or dioritic gneiss. Post-kinematic diorites on Tovqussap nunâ were first described by Berthelsen (1960) who considered that they had formed by replacement of their host rocks. Pillar (1985) interpreted small sheets and segregations of plagioclase-rich rocks in the Nordlandet area as being contemporaneous with granulite facies metamorphism; some or all of these may also belong to the post-kinematic diorites. The post-kinematic diorites are undeformed and unmigmatised and intrude all other Archaean lithologies, except that they are cut by very rare *c.* 5 cm thick pegmatite veins of presumed Late Archaean age. They are also cut by Proterozoic mafic dykes. They form steep or inclined bodies of brown, crumbling, homogeneous rocks (Garde, 1991, fig. 2); their boundaries are up to a few metres wide and gradational with the orthogneiss host rocks, and the marginal parts of their interiors commonly contain partially resorbed country rock xenoliths.

The post-kinematic diorites have very variable modal compositions and mineral textures, but individual bodies appear to be fairly homogeneous. They consist of hypersthene, diopsidic clinopyroxene, hornblende and intermediate plagioclase in variable proportions, besides local coarse-grained biotite (Berthelsen, 1960; Garde, 1991). The rocks typically consist of *c.* 0.5–0.8 cm large, equidimensional hornblende crystals with inclusions of pyroxene and plagioclase, set in a medium-grained plagioclase matrix. Also granular, medium-grained plagioclase-rich rocks occur, and rocks consisting of randomly orientated, subhedral hornblende and pyroxene or biotite, surrounded by plagioclase. A few of the diorites have proto-orbicular textures (*sensu* Leveson, 1966) or contain single (rarely multiple) shelled orbicules (e.g. Berthelsen, 1960). The orbicular textures may indicate crystallisation from a superheated magma (Vernon, 1985). Pyroxene and hornblende in some bodies are mantled by up to *c.* 1 mm thick rims of homogeneous, blue-green ?auto-metamorphic amphibole. However, the post-kinematic diorites do not display retrograde blebby textures of spongy amphibole-quartz intergrowths or sheaves of secondary biotite, such as commonly found in the host grey gneiss. Textures of the post-kinematic diorites are thus partially magmatic, partially ?auto-metamorphic, suggesting that they were emplaced during or after the main retrogressive event.

Age

Garde (1991) reported a conventional U-Pb zircon age of $3017\,^{+10}_{-12}$ Ma (B. T. Hansen, personal communication, 1990) from a diorite plug 3 km east-south-east of Tartorssuaq. The zircon material in the analysed sample 339512 mostly consists of highly irregular crystals and crystal fragments, but there are no signs of partial resorption although the diorite magma was probably superheated when it was emplaced (see below), and the zircon age is practically concordant. The zircons are therefore interpreted as having crystallised from the diorite magma. It might be argued that this *c.* 3000 Ma U-Pb zircon age represents the age of zircons inherited from wall rock orthogneiss, and that the post-kinematic diorites themselves are much younger, perhaps contemporaneous with the Nain plutonic suite (M. Smith, personal communication, 1995) which intruded the Archaean Nain province in Labrador at *c.* 1400 Ma. However, the post-kinematic diorites cannot be young, because Early Proterozoic dykes cut (some of) them.

Also, Rb-Sr isotopic data point to an Archaean age. Strontium isotopic compositions were determined in three samples of post-kinematic diorite (Table 8), likewise collected east-south-east of Tartorssuaq. Figure 80 shows the data plotted in a Rb-Sr isochron diagram together with retrogressed gneiss from the north-western coast of Qugssuk (from Fig. 79b) and a 2954 Ma reference line (the Rb-Sr age of amphibolite facies gneiss at Qugssuk). The post-kinematic diorites plot close to this line, and in spite of their low Rb contents an Archaean age is indicated.

Geochemistry and interpretation

Garde (1991) described the unusual and very variable chemical compositions of the post-kinematic diorites.

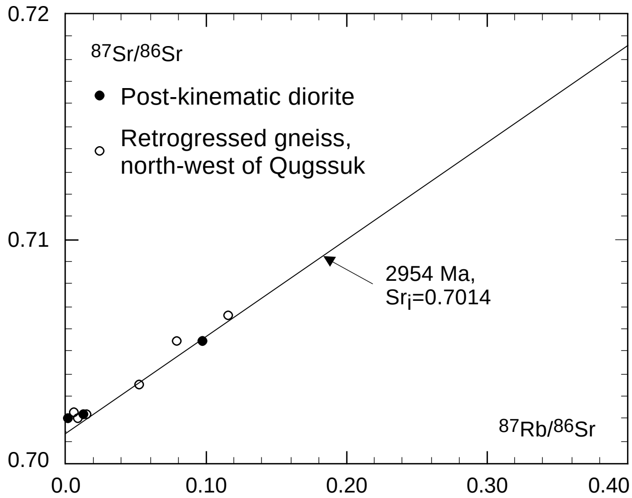

Fig. 80. Rb-Sr diagram showing three samples of post-kinematic diorite, a reference line of 2954 Ma, and samples of retrogressed grey gneiss from the north-west coast of Qugssuk for comparison (see Fig. 79). The samples of post-kinematic diorite are indistinguishable from the grey gneiss, indicating an Archaean age of the diorites.

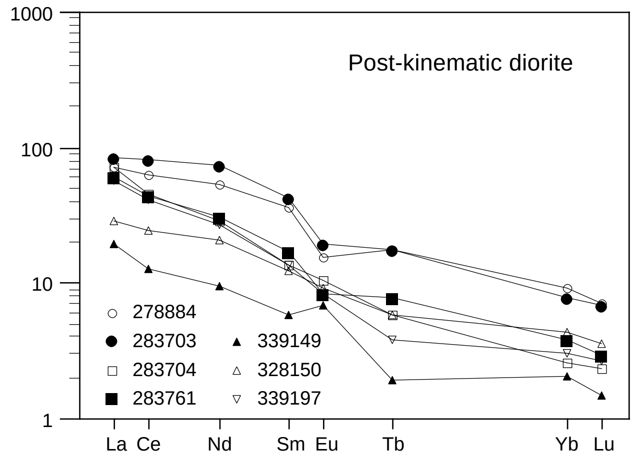

Fig. 81. Chondrite-normalised REE diagram (using normalisation constants by Nakamura, 1974) of post-kinematic diorites from the western part of the Fiskefjord area. Note the extremely variable REE compositions.

The overall intermediate composition of the intrusives is expressed by their silica contents in the range 52.5–58% SiO_2 (Garde, 1991, table 1). They have high

Table 8. Rb-Sr whole rock data for post-kinematic diorite

	Rb ppm	Sr ppm	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	Locality
Post-kinematic diorite					
283703	13.8	409	0.098	0.7055	51°48′17″W, 64°47′38″N
283706	3.6	863	0.012	0.7022	51°47′45″W, 64°47′51″N
283746	0.6	839	0.002	0.7020	51°47′45″W, 64°47′51″N

The samples were collected south-east of Tartorssuaq. See also Fig. 80. Analytical methods as described by Garde *et al.* (1986); the precision of Rb/Sr measurements is within c. 1% (2σ) for samples with more than *c.* 5 ppm Rb, and of $^{87}Sr/^{86}Sr$ measurements better than 0.0002 (2σ).

contents of compatible and mostly very low contents of incompatible elements, and in several samples the concentrations of, e.g. MgO, TiO_2, P_2O_5, Cr and Ni (high contents) and Rb, Nb and Zr (low contents) are comparable to those normally found in ultramafic rocks. REE contents are very variable. Individual samples have widely different REE patterns normalised to chondrite (Fig. 81); collectively the REE contents are higher than in ultramafic rocks, and lower than in (most) dioritic and tonalitic-trondhjemitic grey gneiss in the Fiskefjord area (compare with Figs 22, 58, 62). On the whole, the chemical compositions of the post-kinematic diorites strongly suggest that they were derived from ultrabasic magma with variable contamination or assimilation of continental crustal material.

Garde (1991) noted a general similarity with a group of post-kinematic noritic intrusions some 50 km north of Fiskefjord, with which the post-kinematic diorites may be genetically related. He also noted that the most plagioclase-rich of the post-kinematic diorites are similar in composition to leuconorite and anorthosite dykes in the Nain province, Labrador, described by Wiebe (1979, 1990) to have been intruded as hot liquids above the clinopyroxene-plagioclase cotectic, not as crystal mushes. Garde (1991) concluded that the field relationships, modal mineralogy, textures and chemical compositions of the post-kinematic diorites suggest or are compatible with crystallisation from superheated dioritic magmas. These magmas most likely formed by contamination of hot ultrabasic melts, either with more felsic magma, or by injection of water and assimilation of felsic wall rocks as indicated by field observations and geochemical data reported above.