The Sveconorwegian magmatic and tectonometamorphic evolution of the high-grade Proterozoic Flekkefjord complex, South Norway

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The Precambrian Flekkefjord complex is situated in the core zone of the Sveconorwegian orogenic belt in southwestern Norway, covering the transitional zone between the south Rogaland anorthosite province to the west and the Agder migmatite terrane to the east. The Flekkefjord complex is divided into 3 suites: the banded gneiss suite, the granitic gneiss suite, and the post-kinematic plutonic suite.

The banded gneiss suite contains the oldest rocks with a large spectrum of lithologies of different ages and origin. The oldest part consists of felsic and mafic layers, transformed by bedding transposition into parallelism. They are of unknown age and are of intrusive or extrusive origin, probably serving as the basement for metasedimentary rocks. The latter comprise garnet-sillimanite-cordierite-biotite schists and garnetiferous gneisses, interpreted as metapelites, and quartzite-layered amphibolites, possibly of sedimentary origin. These rocks were deformed (possibly during several phases) and metamorphosed under high-grade conditions during the F_1 phase of deformation, producing intrafolial mesoscopic isoclinal folds with axial plane foliation. The banded gneiss suite was intruded by gabbroic magmas, emplaced as sills and dykes and later deformed and metamorphosed to massive orthopyroxene amphibolites and big-feldspar amphibolites, containing dm-large plagioclase megacrysts. Leucogranitic batholiths were subsequently emplaced into the banded gneisses and later intruded by a 1160 Ma old charnockite.

During a later compressive period, these rocks were deformed into large-scale recumbent folds (F_2), refolded by a later generation of large-scale isoclinal folds (F_3), producing a complicated double-fold pattern on the map-scale. These phases of deformation with accompanying high-grade metamorphism are poorly constrained in the period around 1100±50 Ma BP. The next major intrusive event was the emplacement of batholiths of porphyritic granites with large alkali feldspar phenocrysts (at 1050 Ma), deformed and metamorphosed to augen gneisses during the fourth phase of deformation and high-grade metamorphism. All mafic rocks were metamorphosed under granulite-facies conditions, while only the felsic rocks to the west of Flekkefjord were within the hypersthene-isograd. The deformation resulted in isoclinal folds with N-S trending fold axes and E-dipping axial surfaces, refolding the older isoclinal folds to reclined folds with ENE- to E-plunging fold axes.

At 1000 Ma BP, several granites intruded the region and the Homme granite (990 Ma) obliquely cross-cut and tilted the banded gneiss structure, and was subsequently deformed and metamorphosed in amphibolite-facies. This was the last phase of deformation associated with regional recrystallisation, although the effect was mostly recorded to the east and north. The latest phase of deformation occurred with little recrystallisation, leading to large-scale open to gentle folds with E-W axes and vertical axial planes. Apart from this last phase, the deformation proceeded along N-S fold axes, starting with large-scale recumbent isoclinal folds with amplitudes of tens of kilometres. Their flat-lying axial surfaces became more tilted as the amplitudes of the folds of the younger generations gradually decreased, revealing the piling up of structures in a continent-continent collision in an E-W compressional regime, suggesting an E-dipping subduction zone somewhere to the west.

After the last regional deformation and metamorphism, the post-kinematic plutons intruded the high-grade complex from 980 Ma ago, stabilising the crust as part of the Baltic craton. The granitic magmas generated by partial melting of different rocks in the crustal infrastructure probably above large, underplated magma reservoirs. This mountain chain was subsequently uplifted and eroded and later transected by dolerite dykes between 800 and 600 Ma BP.

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Introduction

The Flekkefjord area is situated in Southwest Norway (Fig. 1) in the southwesternmost part of the Fennoscandian Shield, a subdivision of the Baltic Shield (Holtedahl 1960). Keilhau (1840) was the first geologist to recognise the difference between the Rogaland anorthosite-norite province and the migmatitic gneiss province farther to the east, in Agder. The Flekkefjord area was later considered to be the most typical part of the core zone within the Sveconorwegian orogen. It is dominated by high-grade migmatitic gneisses and amphibolites showing a complicated kinematic interference pattern ascribed to several successive phases of major intrusion and superposed folding. The rocks were ultimately locally deformed and dissected by numerous post-kinematic plutons (Falkum & Petersen 1980, Falkum 1982, 1983, 1985).

The purpose of the present paper is to describe the lithological units and structural evolution and to present a kinematic model to explain the complicated end product. The complexity is ascribed to numerous, successive igneous and

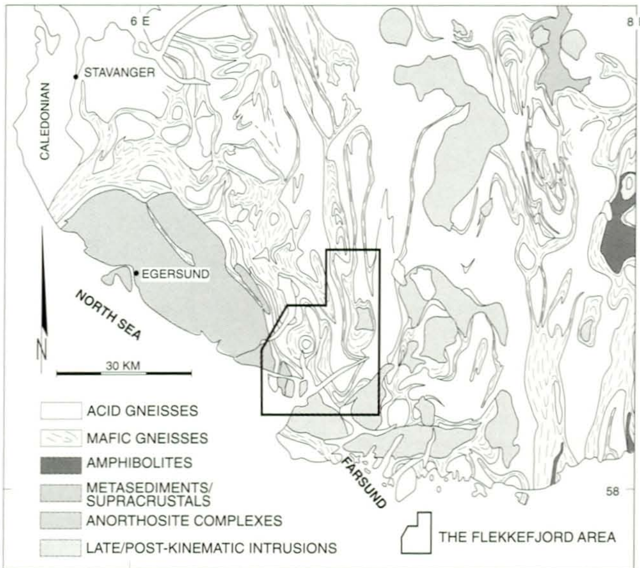

Fig. 1. Simplified geological map of the southernmost tip of Norway with the location of the Flekkefjord area in the border zone between the Egersund anorthosite province to the west and the Agder migmatite terrane to the east. Modified from Falkum and Petersen (1980) & Falkum (1982).

deformational episodes with concurrent regional high-grade metamorphism and partial melting in the mobile core zone during the long-lasting Sveconorwegian orogenic evolution (1250-900 Ma) (Table 1, p. 18). The voluminous intrusions and pervasive deformation have left only a few remnants of pre-Sveconorwegian rocks within this region; these are found especially to the north of the Flekkefjord area (Verschure 1985). The only younger rocks in the area are a few post-orogenic dolerites emplaced between 800 and 600 Ma in the stable Precambrian craton of southern Norway.

Previous work

Southwest Norway is a classic region in the history of geological exploration, beginning in 1820 with the visit of Esmark (1823). He introduced the name *norite* for a series of hypersthene-bearing basic rocks found on the island of Hidra (Hitterøe) and the adjacent mainland. Furthermore, he recognised the white gabbros of the *Egersund anorthosite* massifs. Later, Keilhau (1840) discovered the *Lyngdal hornblende granite* and the *Feda augen gneiss*. Five new minerals were described from the island of Hidra: xenotime (Berzelius 1824) *polycrase* and *malacon* (Scheerer 1843, 1844, 1845), *blomstrandine* (Reusch 1878, Brøgger 1879, 1906) and *kainosite* (Nordenskiöld 1886).

Keilhau (1850) described several new rock types, including a coarse labradorite (i.e. anorthosite) with bluish schiller, and suggested that all contacts were gradational. Dahll (1863) suggested that hornblende-mica schists were older than the subaquatic erupted gneissic granites, which in turn were supposed to be older than Silurian. The gabbros and norites of Esmarks norite formation were considered to be younger than Devonian; and the intrusive hornblende granite of Lindesnes, probably equivalent to the Lyngdal hornblende granite, was regarded to be even younger than the norite formation.

Vogt (1887) opposed the view that all contacts were gradational, considering that contacts were conformable on a large scale, but sharp and in places discordant on a small scale. Furthermore, Vogt (1887, p.18) described an igneous, quartz-rich, alkali feldspar, hypersthene-bearing rock, which he classified as quartz norite. This was 6 years before Holland (1893, 1900) first described charnockite from India. Mafic diabase (dolerite) dykes were described from Hidra (Møhl 1877), and Kjerulf (1883) and Rosenbusch (1883) distinguished two generations of dykes.

The first thorough petrographic description of the anorthosite-charnockite rocks was made by C.F. Kolderup (1897, 1904, 1935) who divided the farsundite into banatite and adamellite, considered to belong to a comagmatic series closely related in time and unaffected by the Caledonian orogeny. Opposing his father, N.H. Kolderup (1929) proposed a Precambrian age, mainly because all rocks were cut by pegmatites of presumed Precambrian age. From the Birkrem area and near the town of Flekkefjord, C. F. Kolderup (1904, 1914) described a layered or banded hypersthene gneiss which he named birkremite, suggesting a genetic relationship to anorthosite.

Granite pegmatites were the focus of interest for the next generation of geologists (Barth 1928, 1931, Andersen 1931, Adamson 1942). Barth (1928) supported the view of Keilhau (1840) and Scheerer (1844) that all rock contacts were gradational. Barth (1935) assumed a co-magmatic origin for the birkremite, anorthosite and farsundite, although he later doubted a magmatic origin for the anorthosite (Barth 1941). Still later, Barth (1945) stated that the anorthosite, farsundite and birkremite were younger than the mixed gneisses.

Howie (1964) analysed hypersthene and augite from a pyroxene granulite (probably farsundite) and concluded that they represented an equilibrium assemblage. Middlemost (1969) also surveyed the farsundite and supported C.F. Kolderups (1897) view that it consists mainly of two different rock types and a minor area of felsic rocks (three main facies). The relationship between these rock types was established by Århus University geologists, who distinguished three different plutons (Falkum et al. 1972). It was shown that the Farsund charnockite intruded the Lyngdal hornblende granite (Falkum et al. 1979), while the Kleivan granite was a third separate intrusion (Falkum & Petersen 1974, Petersen 1977). Furthermore, these plutons were characterised as post-kinematic in relation to the regional deformation and metamorphism, and absolute age determinations showed that they were younger than 1 Ga (Pedersen & Falkum 1975, Petersen & Pedersen 1978, Pasteels et al. 1979).

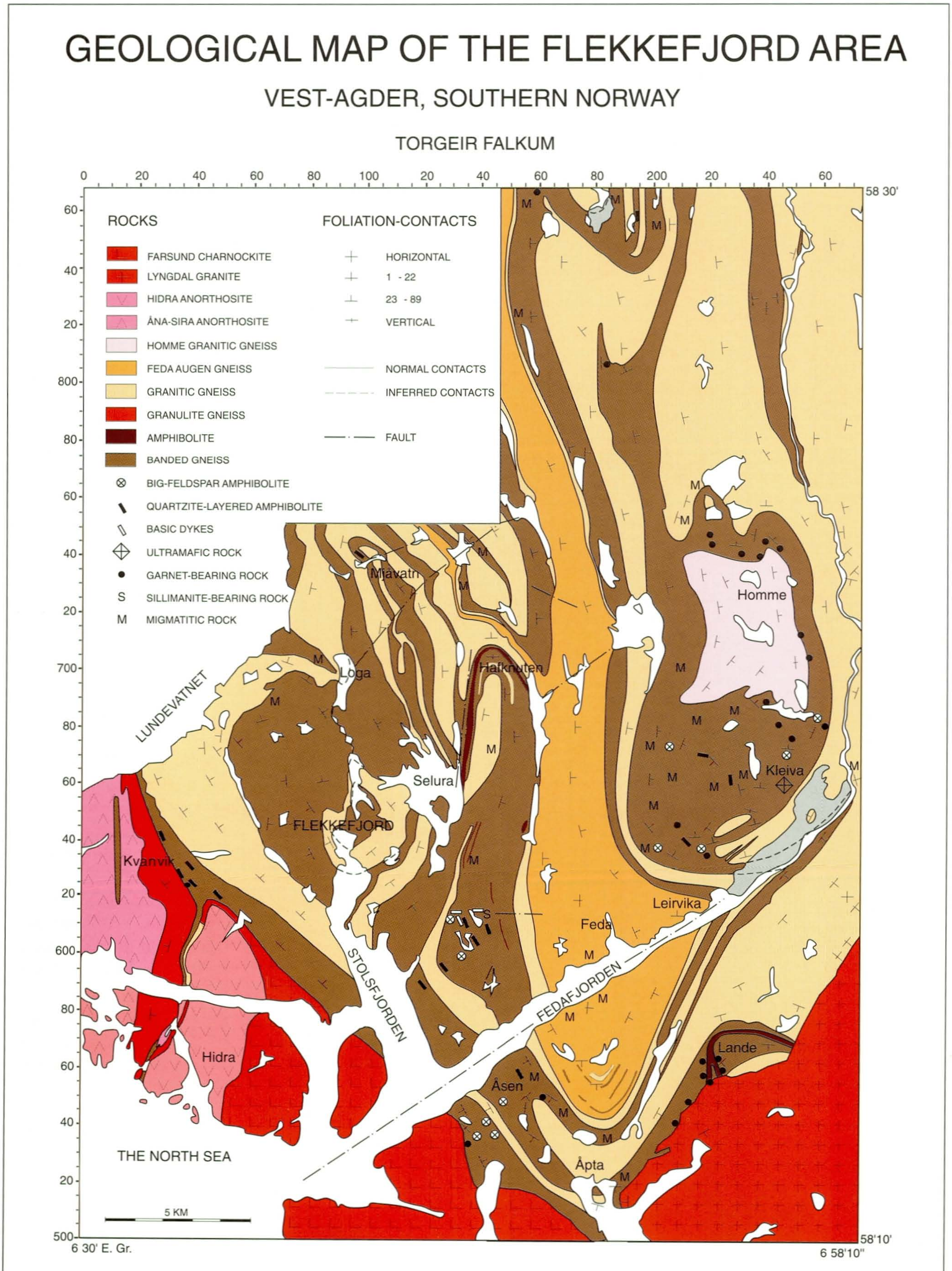

Fig. 2. Geological map of the Flekkefjord area.

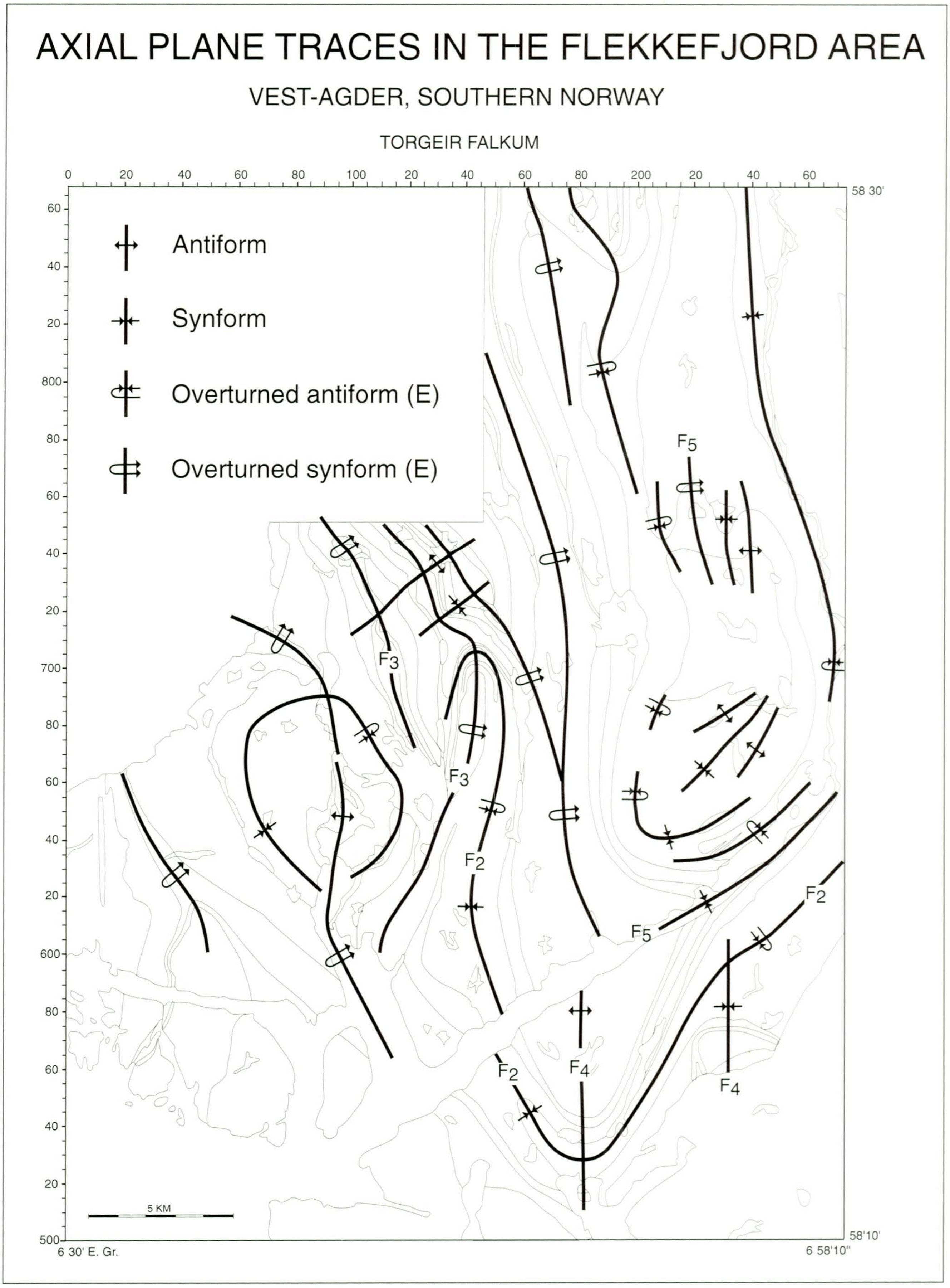

Fig. 2a. Axial plane traces of the refolded Flekkefjord complex.

Subdivision of the Flekkefjord complex

The prevailing view on the Precambrian rock complex of southern Norway when the present investigation was initiated is demonstrated by the following quotations from Barth (1960): «In most other parts granitization is rather complete, rare bands, streaks, or irregularly elongated bodies of amphibolite are the only vestiges of older rocks»; and «The Telemark supracrustals are swimming in a vast sea of granites and granitic gneisses. Within this highly granitized region it has not yet been possible to make out any stratigraphy. Gneisses and granites, often rather monotonous, dominate, and although metasediments are patent in many gneisses, the overwhelming granitization has obliterated the primary structures and thoroughly homogenized the rocks over large areas.» It had only been possible to map some late granites and a few augen gneiss units. The other rocks were regarded as irregular bodies with gradational contacts, as is clearly demonstrated on the maps (Holthedal 1960). Possible large-scale structures could only be discerned by the aid of strike and dip symbols on the map.

Subsequent mapping of a marble/skarn layer in the Tveit area, near Kristiansand, however, revealed the presence of large macrostructures in this highly migmatitic terrane (Falkum 1966a). A large antiform with minor parasitic folds was mapped by dividing the rock units into different lithodemic units, such as gneisses with a predominance of banded, granitic or augen structures. Furthermore, a metasedimentary complex with garnet-sillimanite schist, quartzite and marble/skarn layers was recognised and incorporated into the lithological sequence and mapped as almost continuous layers by detailed tracing of the mega-structures. As a particular unit may appear very different depending on where it is situated within a major fold structure, it is of prime importance to locate the large-scale structures at an early stage and to carry out mapping along the structural trend in relation to the macroscopic structures. This approach is apparently very close to the method of lithostructural mapping suggested by Berthelsen (1960) for high-grade rocks in Greenland.

High-grade migmatite complexes can best be mapped when the structural-dependent lithology is divided into lithostructural units and the kinematic evolution is well understood. Structural interpretation of the mesoscopic structures, combined with an understanding of the map pattern of the lithostructural units, thus revealed an evolution with several successive phases of deformation and intrusive events within the Flekkefjord complex (Fig. 2). Locations on the map are referred to by six numbers; the first three relate to the horizontal scale and the last three to the vertical scale.

Accordingly the Flekkefjord complex has been divided into 3 main lithological suites generally ranging from older to younger rock complexes:

1. **Banded gneiss suite**
2. **Granitic gneiss suite**
3. **Post-kinematic plutonic suite**

Fig. 3. Banded gneiss with alternating felsic and mafic layers dipping to the east along the eastern coast of Selura, looking northeast (loc. 130-650; refers to locality in Fig. 2; 130 on the horizontal scale, 650 vertical scale). The hammer is 40 cm in length.

Each suite comprises several mappable units which are closely equivalent to lithodemic units according to the Norwegian code for stratigraphical terminology (Nystuen 1986, 1989).

The first two suites consist of high-grade metamorphic rocks, either in amphibolite or granulite facies. The last suite comprises several plutons which were emplaced after the last regional metamorphic event.

The _banded gneisses_ consist of sequences of alternating mafic and felsic layers ranging in thickness from less than a centimetre to several metres. Compositionally, the layers range from mafic amphibolites through dark biotite gneisses and further to leucocratic granodioritic and granitic gneisses with a large spectrum of different colours. In addition, this suite also contains units of garnet-sillimanite gneisses, a quartzite-layered amphibolite unit, an amphibolite with plagioclase megacrysts, and massive amphibolites.

The _granitic gneiss suite_ comprises granitic gneisses. Four units can be recognised; (1) The oldest unit consists of relatively homogeneous, faintly foliated, medium-grained, leucocratic biotite granitic gneisses. (2) The next unit is a well foliated augen gneiss with large alkali feldspar megacrysts in the cm-dm range. This is followed by (3) a younger, late-kinematic, discordant granitic biotite gneiss and, finally, (4) a great variety of mafic to felsic, metamorphic, monzonoritic to mangeritic granulite gneisses.

The third suite, named the _post-kinematic plutonic suite_, embraces post-kinematic plutons such as the Lyngdal hornblende granite, the Farsund charnockite, and the Åna-Sira and Hidra anorthosites. The youngest intrusive rocks comprise a few dolerite dykes which intruded the region after it became a stable craton during Late Precambrian times.

Petrography and field relationships

The petrographic description and field relations of the indi-

Fig. 4. A 40 m-high road-cut along E-18 at Leirvika (loc. 225-625) with felsic and mafic layers, ranging in thickness from a few cm to more than a metre. The dark oval area in the upper right is a complicated fold closure where there are mostly mafic layers folded together. Photo taken looking north.

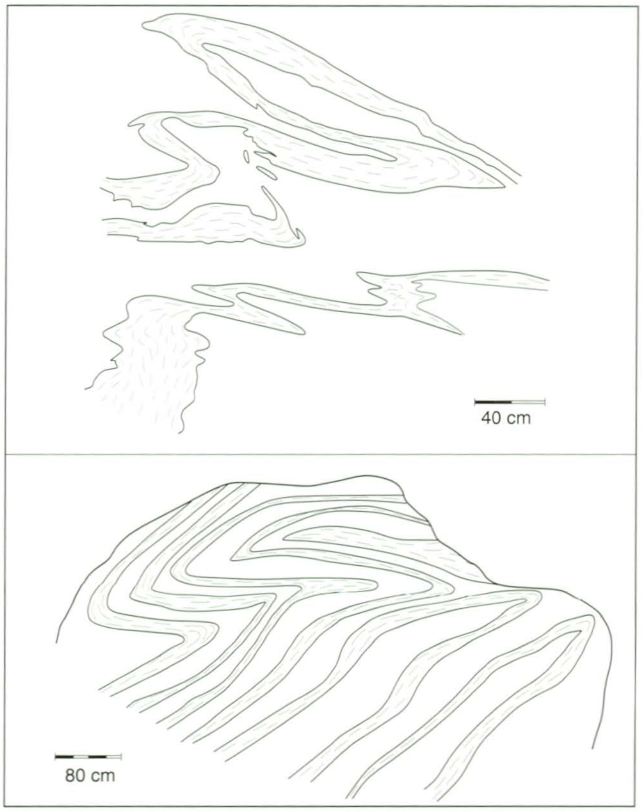

Fig. 5. Sketches drawn from photos of isoclinally folded banded gneiss where the layers are repeated. The upper fold is partly a double fold. In the lower sketch there are several different layers. The mafic layers carry the dashed ornament.

vidual units are dealt with in chronological order. Whereas absolute age determinations are lacking for most of the units, detailed field relationships, combined with an understanding of the structural evolution, commonly reveals the relative ages of the different rock units as well as their role in the long-lasting, repeated processes of deformation and metamorphism.

The banded gneiss suite

The term banded gneiss is used for an inhomogeneous rock consisting of several petrographic rock types confined in discrete layers, giving the outcrop pattern a distinct banded appearance (Fig. 3). The bands are typically in the cm-dm range, although thicker bands occur locally (Fig. 4). When different layers are thicker than 10 to 20 m, they are normally mapped as individual units, if they can be traced over large distances. The most typical rock types in the individual bands range from rather inhomogeneous, felsic gneisses to more mafic gneisses and amphibolites (Falkum 1987a, b).

Generally, the banded gneiss has suffered intense deformation during several phases of folding and shearing. In some outcrops isoclinal folds with one layer repeated up to a dozen times during a single phase of deformation have produced an extremely complicated structural pattern (Fig. 5). Later anatexis added further to the complexity, resulting in a variety of migmatitic structures.

Felsic layers

The most common banded gneiss units comprise a layered rock-type with alternating felsic and mafic bands or layers. The majority of felsic layers are gneisses of granitic to granodioritic or more rarely tonalitic composition.

In the field, the felsic layers show a considerable variation in composition, texture and colour and small-scale variations result in rather inhomogeneous leucocratic gneisses. Most commonly they show a great variation with streaks and pods of different composition as well as a large variation in grain size and colour both along and across the foliation within the individual layers. Strong deformation in connection with the formation of isoclinal folds, resulted in a distinct contact-parallel foliation, contributing to the well-layered appearance of the rock. The banding is accentuated within some of the felsic layers by the presence of 1-2 mm-thin, slightly darker bands consisting mostly of biotite and dark feldspars. These bands are generally extremely thin and are in places strongly disturbed by folding and pinch-and-swell structures where the dark layers wedge out within the leucogneissic matrix. Due to the extreme plastic deformation combined with local anatexis, a wide variety of migmatite structures are present in these rocks.

Apart from this compositional layering, a faint banding due to colour and/or grain size variation is also discernible. The colour variation is mainly caused by feldspars with different colours, partly due to different amounts of inclusions, either foreign or exsolved minerals.

Fine-grained felsic layers locally grade into pegmatitic rocks, which in certain instances become deformed and in rare cases cross-cut the foliation and layering within the banded gneiss sequence, a further indication of local mobilisation within discrete layers.

The microtexture of the light-coloured gneisses is normally hemi-granoblastic with polygonal grain boundaries. Quartz and feldspar make up >90% of the rock. The anhedral quartz grains have normally strong undulatory extinction and usually contain numerous fluid inclusions situated away from grain boundaries and internal cracks.

The alkali feldspar is generally orthoclase, commonly with undulatory extinction and only rarely with an extremely faint cross-hatching, only developed within small areas in a few grains. Plagioclase (An_{22-30}) with well-developed pericline twins, is in places partly or even totally altered, mostly to sericite.

The felsic layers in the westernmost banded gneisses generally have a granulite-facies mineral assemblage with small amounts of orthopyroxene (hypersthene), clinopyroxene (augite) and hornblende, in places considerably altered. Dark-brown biotite may also be present. In the eastern and central parts of the area, an amphibolite-facies mineral assemblage prevails in the leucocratic rocks with a Ti-rich pale brown to red biotite as the dominant mafic mineral with subsidiary hornblende and sphene. Zircon and apatite are common accessory minerals in both types, together with extremely rare opaque minerals, mostly ilmenite and magnetite, in places altered to hematite. Secondary minerals comprise zoisite, chlorite and white mica.

Mafic layers

A great variation in grain size, colour and texture is also seen in the mafic layers. They vary from the mm-scale to several metres in thickness and are mostly medium- to fine-grained, although rather coarse-grained layers with large internal grain-size variation occur throughout the area. The dominance of pyroxene, amphibole and biotite results in a dark, homogeneous, eugranoblastic texture, where only more detailed inspection reveals a foliated rock with nematoblastic or even lepidoblastic texture. The layering is commonly strongly disturbed by intense folding and pinch-and-swell structures and many mafic layers wedge out into felsic-dominated domains.

Plagioclase (An_{40-50}) is the dominant felsic mineral, normally comprising more than 50% of the total rock. Orthoclase and quartz are found locally, but are normally absent. Hypersthene and augite are the dominant mafic minerals in the granulite-facies rocks with minor amounts of hornblende and biotite.

Garnet-sillimanite schists and gneisses

The most common type of sillimanite-bearing rock is a leucocratic, quartz-alkali feldspar-dominated garnetiferous gneiss with thin bands rich in almandine-garnet and sillimanite. Microscopically, these zones also contain cordierite and zircon together with a subordinate amount of apatite, green spinel, plagioclase and biotite. Opaque minerals are rarely present in these rocks. A large variation in grain size is found in this rock-type with hemi-granoblastic texture, although the sillimanite-rich zones can be nematoblastic. Quartz is normally well rounded and contains fluid inclusions. The amount of fluid inclusions in these quartz grains is considerably less compared with the amount in quartz from the felsic layers in the banded gneisses. Microcline with cross-hatched twinning is the most common alkali feldspar, but poorly twinned grains with a tendency towards mesoperthitic texture also exist. Most of the few, small zircons seem to be well rounded.

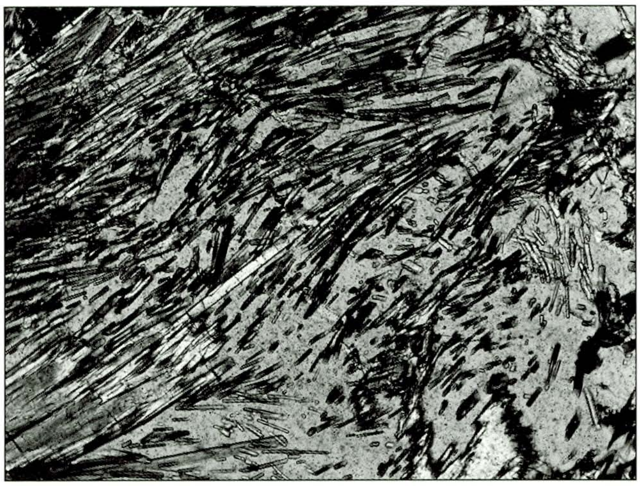

Fig. 6. Photomicrograph of a sillimanite-rich zone in the garnet-sillimanite-cordierite-biotite schist where the sillimanite penetrates cordierite, alkali feldspar, plagioclase, and biotite. The photo is 1 mm across.

A second type is essentially a garnet-sillimanite-cordierite-biotite schist containing more than 50% of red-brown Ti-rich biotite. Almandine-garnet and commonly fibrolitic sillimanite also constitute a major part of the schist (Fig. 6). In addition, cordierite, mesoperthitic alkali feldspar, sericitised plagioclase, apatite, green spinel, a few small rounded zircons, and opaque minerals are found microscopically. Quartz is normally absent, but has been observed in some thin-sections. Anomalous blue zoisite, white mica and myrmekite seem to be of secondary origin. Intense deformation of these garnet-sillimanite-cordierite rocks resulted in very complicated folds. They also occur as isolated lenses, which can be followed for 50 to 100 metres along strike. However, the lenses appear repeatedly at the same stratigraphic level, so they would appear to be stratabound. The question is whether these rocks represent original clay-rich sediments or if they attained their peraluminous character by metasomatism, or represent residual rocks after removal of granitic melts formed by partial melting.

Quartzite-layered amphibolite

Within the mafic layers in the banded gneiss there are small zones with alternating cm-thick layers of orthopyroxene amphibolite (pyribolite) and quartzite (Fig. 7). The quartzite layers could originally have been intrusive quartz veins, but they are never seen to cross-cut the foliation. The possibility of a sedimentary origin therefore exists.

The quartzite layers consist of strongly elongated quartz grains with thin inclusions of mostly cloudy feldspars, all lying parallel to the layering. The boundary between the quartzite layer and the amphibolite consists of a coherent

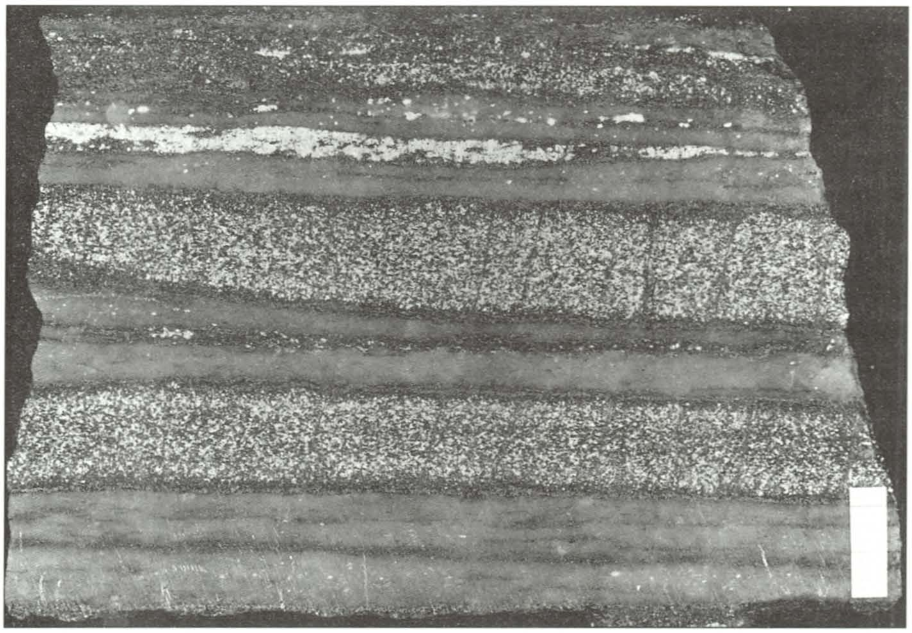

Fig. 7. Cut surface of a quartzite-layered amphibolite with internal layering within the quartzite layers. The white dots within the amphibolite layers are individual plagioclase grains. The scale, lower right, measures 12 x 4 mm.

thin band of faintly red orthopyroxene (hypersthene) in places rimmed by clinopyroxene (augite). This thin layer of pyroxenes is followed by a zone with plagioclase, twice as thick as the pyroxene layer. Farther away from the quartzite the plagioclase zone grades into the normal orthopyroxene amphibolite, which consists of orthopyroxene, clinopyroxene, plagioclase, opaque minerals, and hornblende which in some cases seems to replace the pyroxenes. In certain zones the plagioclase may be considerably altered to sericite and elsewhere the pyroxenes are strongly altered and fresh biotite is abundant.

Big-feldspar amphibolites

This deformed and metamorphosed unit with plagioclase megacrysts and anorthosite fragments up to several dm in size (Fig. 8) is described in detail elsewhere (ms. in prep.).

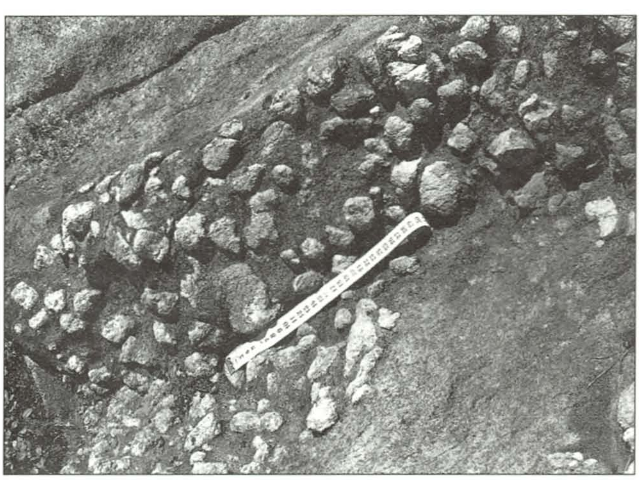

Fig. 8. The big-feldspar amphibolite with plagioclase megacrysts and anorthosite fragments embedded in a hypersthene amphibolite (pyribolite) (loc. 260-685). The cm-measure is 30 cm.

It is concluded that these rocks represent several sills and/or dykes, the majority of which intruded roughly along the same horizon. Although the big-feldspar amphibolites are strongly deformed and metamorphosed, they are comparable to unmetamorphosed Gardar mafic dykes from Southwest Greenland with plagioclase megacrysts and/or blocks of anorthosite, named the big felspar dykes by Bridgwater (1967) and Bridgwater & Harry (1968). They concluded that the megacrysts had grown in the rock in which they are situated, so there existed «a cognate rather than an accidental relationship between inclusions and the magmas in which they became incorporated.»

In the Flekkefjord area these big-feldspar amphibolites intruded before 1160 Ma BP, suggesting that there were two separate periods of intrusion of anorthosite-kindred rocks as the emplacement of the large Egersund anorthosite massifs to the west took place around 930 Ma ago.

Massive orthopyroxene amphibolites

These orthopyroxene amphibolites comprise homogeneous, massive, dark layers in the banded gneiss, varying in thickness from a few tens of metres up to between two and three hundred metres. In well-exposed areas they can be followed for several kilometres along strike, but all of them die out ultimately.

The amphibolites normally lie concordantly within the banded gneiss, but one of them cross-cuts the foliation and lithological layering, revealing an intrusive origin. This occurs within the synformal structure at Lande (225-560) and the intrusion post-dates the F_1 structures. Together with the synform the amphibolite is folded around the large, isoclinal F_3 folds. None of these amphibolites have been observed in the granitic gneisses which are folded by the F_2 folds. This suggests that the massive amphibolites intruded between the first and second phases of regional deformation and before intrusion of the granitic gneisses.

The orthopyroxene amphibolite is macroscopically a rather homogeneous, coarse-grained rock with a strongly foliated, nematoblastic texture with orthopyroxene (hypersthene), clinopyroxene (augite) and hornblende as the major mafic minerals together with plagioclase (An_{40-50}). Faint banding on a cm-scale, due to different amounts of mafic minerals, occurs locally. It is not clear whether this represents original igneous layering or is a result of intense ductile deformation and metamorphic differentiation.

Some of the thicker layers, and especially the Lande amphibolite, have a 3- to 4 m-wide border zone which is very homogeneous and fine grained with almost no macroscopically visible foliation. This zone, which now contains a totally recrystallised metamorphic mineral assemblage, probably represents a chilled margin.

Ultramafic rocks

A small body of ultramafic composition is found at one locality within the Flekkefjord area. It is a semi-circular body approximately 50 m in diameter which is situated at the farm Kleiva (250-665). It is a coarse-grained, dark-green pyroxenite, with large crystals of green clinopyroxene as the dominant mineral together with smaller and fewer grains of orthopyroxene and a few, small, altered grains of olivine.

This body represents a stock-like intrusion which has been strongly deformed. Although the age is uncertain, it is considered to have intruded at a relatively early stage in the deformational evolution.

The granitic gneiss suite

This group of rocks comprises several types of granitoid plutonic intrusions, all of which have suffered at least one phase of deformation and regional metamorphism. They vary in size from small plugs or sills to batholith-sized bodies covering several hundreds of km^2.

Granitic gneisses

The granitic gneisses can be followed on the map as major lithological units throughout the Flekkefjord area. In most places they are medium- to coarse-grained, homogeneous gneisses with a granitic composition (Fig. 9). The colour is commonly pale grey on slightly weathered surfaces, whereas it becomes more dark grey in completely fresh outcrops. In the west it tends to be more dark green or brown, whereas in the east and north it is grey to pink. It may be impossible to detect any foliation in the most leucocratic zones within the gneiss. In other places a faint foliation can be observed related to banding on a cm-scale due to variations in mafic mineral content or grain size.

Towards the banded gneiss, the granitic gneiss may contain more mafic and banded zones with biotite-rich layers consisting of 1-2 mm-thin biotite layers spaced 1 to 2 cm apart (Fig. 10). This layering may have a constant strike, but in most areas it is strongly deformed with the thin biotite layers folded into several isoclinal fold-hinges. Still closer to the boundaries, the dark layers may reach up to several cm in thickness, and if the rock is coarse-grained, oval-shaped alkali feldspar, up to one cm in length, can be common and the rock then has the character of a leucocratic augen gneiss. These zones could represent strongly deformed inclusions of banded gneiss which have been sheared and flattened to an extreme degree.

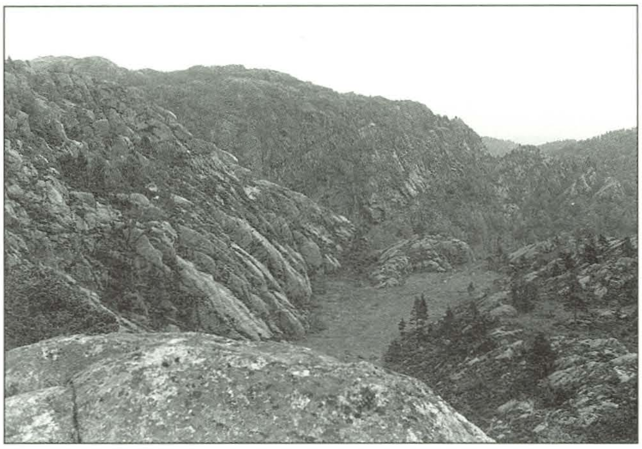

Fig. 9. Homogeneous granitic gneiss from the fold core of the Hafknuten reclined fold (loc. 145-690) looking south. The east-dipping foliation can be seen where there are slightly more biotite-rich layers.

Quartz-rich veins, together with pegmatites and aplites, are ubiquitous, either as straight foliation-parallel layers or as folded and pinched-off lenses or veins. Pegmatites commonly grade into normal granitic gneiss.

Microscopically, the granitic gneisses have hemi-granoblastic textures, although the most biotite-rich zones show lepidoblastic textures which may be detected macroscopically as a foliation. Quartz and mesoperthitic alkali feldspar totally dominate the rock, although a certain amount of plagioclase (An_{25-35}) may be present in some areas, and regionally the plagioclase content shows a large variation. The com-

Fig. 10. Indistinct biotite-rich zones occur from dm- to several tens of metres in thickness in the granitic gneisses. They are almost always intensely deformed, and commonly show isoclinal folds. Photo taken east.

position of the gneisses therefore varies from alkali feldspar granitic through granitic to granodioritic.

Yellow to red-brown biotite is always present in minor amounts (2-10%) and it commonly replaces orthopyroxene and clinopyroxene which are usually altered. Hornblende is rare, but can be found in the northeastern part of the area together with biotite and well cross-hatched microcline in a typically pink-coloured gneiss. Large zircons are common.

The granitic composition and general homogeneity over vast areas point to an intrusive origin. Since these gneisses have undergone several phases of deformation, the boundaries are overall conformable with the banded gneiss. However, on the mesoscopic scale, the border of the granitic gneiss is locally seen to run parallel with the foliation in the banded gneiss, but in a few places makes a sharp, right-angle turn into the neighbouring rock. This step of up to half a metre is a cross-cutting relationship which is a pre-deformation event, since the contact-parallel foliation cuts straight across this discordant border. Although these step features may have been created during an early phase of deformation, it seems more likely that they represent a primary cross-cutting intrusive relationship.

Another observation pointing to an intrusive origin is the local presence of many banded gneiss inclusions with different orientations. This could also be due to local folding, but no such structure could be detected around the inclusions and the foliation within the granitic gneiss is normally of a consistent trend over the whole outcrop. Some of the inclusions have straight edges, but most of them are rounded with ovoid or lenticular shape. Foliation and lineation trend undeflected through these inclusions, usually obliquely to their elongation and internal layering, clearly showing that they were in place before the regional deformation and metamorphism. Several of the inclusions grade continuously into the granitic gneiss, whereas others, generally the most mafic, have a biotite-rich border zone, possibly a result of reactions with the magma.

Augen gneisses

Another type of felsic gneiss is a texturally distinct rock with large alkali feldspar megacrysts deformed to oval augens and classified as augen gneiss (Fig. 11). This group of easily recognisable rocks constitutes a major lithology in southern Norway, formed as porphyritic granites, commonly of batholithic size and later deformed and metamorphosed, resulting in the typical texture with ovoid-shaped feldspar with encircling mafic minerals. One of these bodies, situated in the area immediately to the southeast of the Flekkefjord area, has been treated in detail by Petersen (1977).

The main unit which makes up the spine on the Flekkefjord map-sheet is the Feda augen gneiss, discovered by Keilhau (1840). It is a spectacular grey to bluish-grey rock with up to 15 to 20 cm large alkali feldspar megacrysts (orthoclase). Reconnaissance mapping revealed that this elongate, ovoid, doubly-folded body is discordant to the country

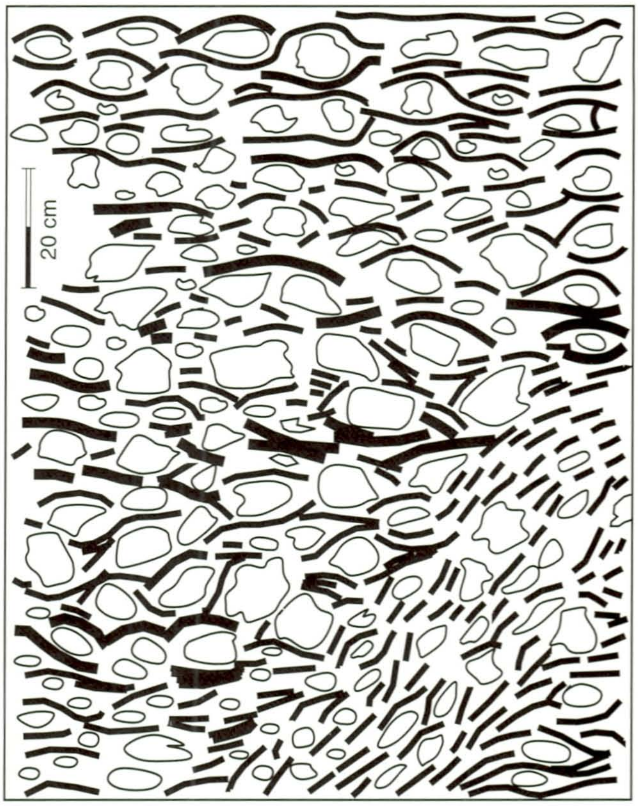

Fig. 11. Sketch of the Feda augen gneiss showing the different sizes and shapes of K-feldspars and the biotite foliation curving around the megacrysts. Local shear zones (lower right) are ubiquitous.

rocks on a regional scale. This was at first interpreted as an erosional unconformity (Falkum 1966b), but later structural studies have revealed that it was the result of a transgressive, discordant intrusive contact (Falkum 1985).

The original porphyritic granite, with large grain-size variations, has been strongly deformed and metamorphosed under amphibolite-facies conditions resulting in an inhomogeneous rock. The texture is hemi-granoblastic with an uneven distribution of quartz, alkali feldspar (one measurement, 16 x 6 cm), and plagioclase (An_{25-30}) together with clusters of biotite, amphibole, clinopyroxene, apatite, zircon and opaque minerals. Inclusions of different compositions are common and variably assimilated. Together with the host augen gneisses they were later intensely deformed during several phases of deformation, and this feature adds considerably to the impression of an extremely inhomogeneous body.

Some of the inclusions of granitic or banded gneiss occur as layers up to 50-60 m in thickness. Only a few of the largest inclusions are drawn on the map. They are numerous in the field, especially along the eastern contact and southwards to the main hinge zone of the Feda fold (see map). The southernmost layer of biotite-rich, banded gneiss inclusions, which is folded around the Feda hinge-zone, contains garnets, possibly suggesting a higher pressure within the fold core than in the limbs.

The fold hinge-zone and peripheral parts of the body are extremely strongly deformed and recrystallised, resulting in a true augen texture, whereas the central parts are considerably less deformed with almost euhedral megacrysts which may be randomly orientated. These orthoclases are commonly filled with contact-parallel inclusions and Carlsbad twins are ubiquitous, suggesting that the megacrysts nucleated and grew during the magmatic stage and developed their oval form during metamorphic recrystallisation in connection with the post-emplacement phases of deformation and regional metamorphism.

Pegmatites and aplites are ubiquitous and rare mafic dykes and sills can be recognised, although they have been deformed and metamorphosed. Chemical analyses show that these mafic rocks are P- and K-rich, and probably represent potassium-rich lamprophyres.

The original magma intruded discordantly through the banded and granitic gneiss units after the third phase of deformation and crystallised as a porphyritic granite, subsequently deformed and recrystallised during the last three phases of deformation and metamorphosed under high-grade conditions. As it has been radiometrically dated, it constitutes an important time marker in the structural evolution, as well as providing a clue to the physical conditions during the metamorphic events. The Feda augen gneiss was studied in detail by Bingen (1989) and dated by Bingen et al. (1990, 1993, 1996) to 1040 ± 44 Ma (Rb-Sr) and later to 1051 +2/-4 Ma (U-Pb)(Bingen & van Breemen 1998).

Late-kinematic discordant granitic gneisses

There is only one major, deformed and metamorphosed pluton with a high-angle cross-cutting contact relationship on the Flekkefjord map-sheet, but there are several of these late-kinematic plutons immediately to the east. The main body is a deformed biotite granite which intruded before the last regional metamorphic event, named the Homme granitic gneiss, described in detail elsewhere (Balling Rasmussen & Falkum 1975, Falkum 1976a,b, Falkum & Rose-Hansen 1978, Falkum & Pedersen 1979, Falkum et al. 1987).

Radiometric datings of this late-kinematic intrusion and the oldest post-kinematic pluton constrain the last regional deformation and metamorphism which occurred during the waning stages of the Sveconorwegian orogeny. A Rb-Sr age determination of the Homme granitic gneiss, interpreted as the age of intrusion, yielded 997 ± 14 Ma with an initial Sr ratio of 0.7043 (Pedersen & Falkum 1978) later revised to 990 ± 16 Ma (S. Pedersen pers. comm. 1997). The oldest post-kinematic intrusion is the Holum granite, 30 km to the southeast of the Flekkefjord area, which gave a Rb-Sr age of 980 ± 33 Ma and an initial ratio of 0.7045 (Wilson et al. 1977).

Monzonoritic-mangeritic granulite gneisses

The metamorphic grade generally increases towards the west and the westernmost zone, in contact with the Åna-Sira anorthosite, contains several types of dark bluish to brown granulite-facies gneisses. The main rock-type is a mangeritic to monzonoritic (jotunitic) gneiss, and quartz-free leuconoritic gneisses can also be found. There is, however, a large variation in composition from charnockitic through opdalitic to enderbitic gneisses, although the majority of rocks are either quartz-poor or quartz-free. The most common type has a hemi-granoblastic texture with polygonal grain boundaries of plagioclase (An_{25-35}), mesoperthite, orthopyroxene, clinopyroxene, biotite, apatite, and opaque minerals. In the most deformed granulites, the internal grain boundaries of quartz and feldspar are normally interlobate or even amoeboid and in extreme situations the mesoscopic foliation is dominated by elongated and platy mineral aggregates, giving a wavy, tabular appearance.

Most of these rocks are intrusive and there are several bodies which have been deformed and metamorphosed together. These gneisses are always well foliated with platy or augen-shaped feldspars in between the mafic minerals which generally cluster in thin, parallel layers or aggregates. The foliation commonly becomes weaker towards the anorthosite contact and in some places disappears completely. In other instances it is contact-parallel, but in many cases it runs oblique to the contact. The anorthosite is intrusive into the monzonoritic/mangeritic gneiss, as can be seen from the curving contact on the map, contrasting with the smooth eastern contact of the monzonoritic/mangeritic gneiss. These gneisses are apparently older than similar rocks to the north, as they have suffered several phases of deformation and regional metamorphism.

The post-kinematic plutonic suite

The last phase of deformation and regional metamorphism terminated before the intrusion of the Holum granite at 980 Ma (Wilson et al. 1977). Accordingly, this pluton and all the younger intrusions are classified as post-kinematic in relation to the orogenic movements, but are included in the last main phase of the orogenesis. The rock units composing this group are, among others, the anorthosite-norite-mangerite suite to the west, thoroughly investigated by research groups from the universities in Brussels, Liege, Århus and Bergen (Michot & Michot 1969, Demaiffe & Michot 1985, Duchesne et al. 1985, Wilson et al. 1996, Duchesne & Wilmart 1997, Robins et al. 1997). The *Farsund charnockite* in the southern part of the Flekkefjord map also belongs to this suite (Falkum et al. 1979).

There are also many undeformed biotite and/or hornblende granitic plutons unevenly distributed in South Norway and southwestern Sweden. In the Flekkefjord area, an example of one of these plutons is the *Lyngdal hornblende granite*, which discordantly intruded the granitic and banded gneiss units in the southeastern corner of the Flekkefjord map. A U-Pb zircon date gave an age of 950 Ma (Pasteels et al. 1979), within the range of the Rb-Sr determination, 932 ± 38 Ma with a low initial Sr ratio of 0.7054 (Pedersen & Falkum 1975). The Rb-Sr ages are apparently generally younger than the U-Pb ages, as shown in particular for the Farsund charnockite with a Rb-Sr age of 874 + 48 Ma (S. Pedersen pers. comm. 1997) compared with a U-Pb age between 920 and 940 Ma (Pasteels et al. 1979).

The *Hidra anorthosite* has the youngest U-Pb age at 930 Ma and a Rb-Sr age of 892 ± 25 Ma (Pasteels et al. 1979, Weis & Demaiffe 1983). The *Åna-Sira anorthosite* was probably emplaced during the same time span. These plutons are described from northwest to southeast.

The Åna-Sira anorthosite

The Egersund anorthosite province is built up of large semicircular massifs, situated near the southwest coast of southern Rogaland. Interpretation based on aeromagnetic surveys suggests that the major part of the province is hidden below the sea in direct continuation from the exposed massifs (Holtedahl & Sellevoll 1971, Sellevoll & Aalstad 1972, Sigmond 1992). Mapping in the present project covered the southeastern part of this province, but no detailed study was performed as this had already been carried out by others (J. Michot & P. Michot 1969, Demaiffe & Michot 1985, Duchesne et al. 1985 and Duchesne & Michot 1987).

When fresh, the coarse-grained anorthosite is dark, commonly with a bluish schiller in the plagioclase (An_{45-55}), whereas the weathering colour is brown or grey to white. Along the boundary many large (5-20 cm) plagioclases are surrounded by small plagioclase grains; macroscopic orthopyroxene is rare. The zones with the small grains are probably a result of post-crystallisation deformation (granulation). In those cases where some of the larger crystals are elongated, they tend to be parallel to the contact.

One conspicuous feature is the presence of numerous inclusions of felsic and more rarely mafic rocks. The largest inclusion is shown on the map to the west of Kvanvik (025-630) and is several hundred metres wide and over 4 km from north to south. N-S-trending shear zones are commonly found in connection with these inclusions, and some E-W-trending breccia zones have been encountered. West of Kvanvik (025-630) the main road follows one of these E-W breccia zones, and close to it is a small (20 m) semi-circular body of ilmenite-norite.

The transgressive boundary shows that the anorthosite intruded and partly recrystallised the monzonoritic/mangeritic gneiss by contact metamorphism. Whether the intrusion was a magma or crystal mush is difficult to discern, but as the body continued to be deformed after emplacement it could have been a slow-moving crystal mush.

The Hidra anorthosite

The southeasternmost anorthosite is a small pluton, situated as a 3 km-wide lens in the middle of the island of Hidra, extending northwards on the mainland for another 3 km and wedging out into the core of the Kvanvik antiform (050-615). For a more detailed description, reference should be made to Demaiffe et al. (1973), Demaiffe (1977), Demaiffe & Hertogen (1981), Weis & Demaiffe (1983) and Demaiffe & Michot (1985).

The sharp eastern contact with the Farsund charnockite dips steeply to the east. In several places and especially along the eastern boundary, the content of mafic minerals, mostly orthopyroxene, increases to more than 20% in the border zone, thus merging into the leuconorite field. At one locality (061-600) the leuconorite has a 3-4 m monzonoritic layer and further a thin banded gneiss zone before the charnockite is encountered. This monzonorite (jotunite) has been considered to represent the parental magma of this pluton (Demaiffe & Hertogen 1981).

Along its northwestern border, the leuconorite has a relatively sharp contact, within a cm or two, with a charnockitic gneiss. The colour of this gneiss varies from pale grey to pale brown towards the leuconorite, related to an increasing proportion of mafic minerals. The western contact on Hidra is a complicated agmatitic zone; for a more detailed description reference should be made to Demaiffe et al. (1973), Demaiffe (1977), Demaiffe & Hertogen (1981), Weis & Demaiffe (1983) and Demaiffe & Michot (1985) .

The Hidra anorthosite/leuconorite is generally finer-grained than the Åna-Sira anorthosite. Apart from a few zones within the central part of the Hidra body with large plagioclase crystals (An_{45-55}), the average grain size is in the range of 1 to 5 cm, and in many cases less than 1 cm (An_{40-50}). In several places, more than 10% of orthopyroxene is present and the texture becomes subophitic. This undisturbed texture, combined with the sharp boundary, suggests that magmatic crystallisation occurred in the present position without much displacive movement in the post-crystallisation stage.

The Farsund charnockite

The Farsund charnockite, which is the charnockitic part of the farsundite (Kolderup 1904), is one of the latest, major, post-kinematic intrusive plutons in Southwest Norway. For a more detailed description, reference should be made to Falkum et al. (1974, 1979).

The Lyngdal hornblende granite

This pluton was earlier also considered to be part of the farsundite, but more detailed investigation showed that it was a separate intrusion, named the Lyngdal hornblende granite (Falkum et al. 1979). Another pluton within the farsundite complex was also distinguished as an individual pluton to the northeast and named the Kleivan granite (Falkum & Petersen 1974, Petersen 1980a, b). It is considered to have evolved in a separate magma chamber.

Both mineralogically and geochemically the Lyngdal hornblende granite is different from the Farsund charnockite and the Kleivan granite. The Lyngdal granitic magma intruded the metamorphic country rock complex discordantly and the cross-cutting contact relationship between the Lyngdal hornblende granite and the country rocks is clearly demonstrated in the southeastern corner of the Flekkefjord map-sheet (Falkum et al. 1979).

The late dykes

The few late mafic dykes that have been observed are fine-grained, cross-cutting dolerites usually around a metre in thickness. Most of them are vertical and strike E-W, but a few strike roughly N-S. In most areas only a single dyke has been found; but in two areas on the Flekkefjord map there are up to a dozen dykes within a small area. In the Tjersland area (135-615), five dykes have been observed, but it has not been possible to follow them for more than a score of metres. Most of the dykes have been found in the westernmost monzonorite/mangerite gneiss unit and they apparently become more abundant in the anorthosite massifs to the west of the Flekkefjord map-sheet where they have been dated to the period between 800 and 630 Ma (Storetvedt & Gidskehaug 1968, Versteve 1975, Sundvoll 1987).

Structural analysis of the Flekkefjord area

Mapping in a high-grade metamorphic terrain is difficult unless the structural evolution within the region can be established. As the multiple phases of deformation proceeded under highly ductile conditions in the Flekkefjord area, similar styled mesoscopic folds developed during the different phases of deformation. Refolding of macrostructures must consequently be distinguished and mapped in order to build up the chronological sequence of structural events. In this connection it is important that several dated plutons and batholiths intruded the complex between, or during, the separate dynamic phases, giving a possibility for reconstruction of the time-sequence for the separate phases of deformation.

Development of foliations and lineations

In order to understand the structural evolution, the mesoscopic structures are defined and described, followed by a geometrical analysis of these structures associated with the macroscopic fold-structures with the aim of revealing the time-sequence of both structural and intrusive events during the evolution of the Flekkefjord area.

Foliation is considered as any mesoscopically penetrative parallel alignment of planar fabric elements (s-surface) which is observed in the metamorphic rocks. The most common type is a preferred orientation of platy and/or prismatic minerals, like biotite and amphibole, and the foliation becomes weaker as the metamorphic grade increases because biotite and amphibole are replaced by pyroxenes.

Another type is the compositional layering which is well developed in felsic rocks, and to some extent also in mafic rocks. The mineral foliation is almost always parallel to this compositional layering, as is the case with the third type of foliation, the colour banding.

The last type of foliation occurs in the augen gneisses where large lenticular megacrysts are orientated subparallel to the prevailing mineral foliation and the mafic minerals (mostly biotite and hornblende) curve around the ovoid megacrysts (Fig. 11). When these ovoid feldspars are larger than approximately one cm in length, bending of the mafic minerals creates a considerable curvature, resulting in a wavy foliation.

In conclusion, it can be stated that practically all layered metamorphic rocks in the Flekkefjord area characteristically

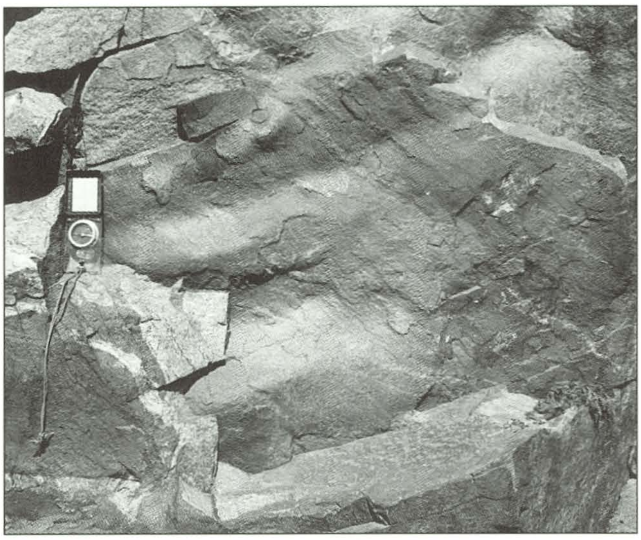

Fig. 12. Foliation surface with a crenulation (F_4) plunging to the right, refolded by gentle folds (F_6) with fold axes plunging to the left. From the banded gneiss to the west of Åpta (loc. 165-125), looking northeast.

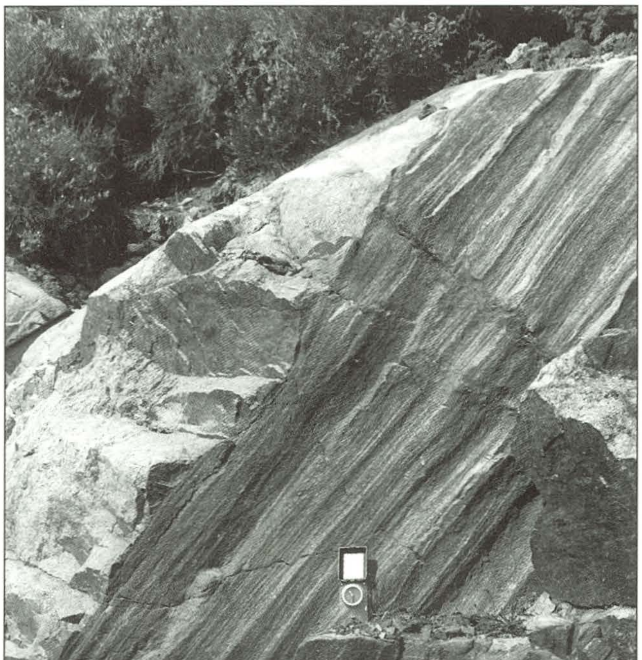

Fig. 13. South-plunging crenulation (F_4) in the banded gneiss 500 m to the west of the axial plane trace of the Åpta antiform (loc. 175-525), looking west.

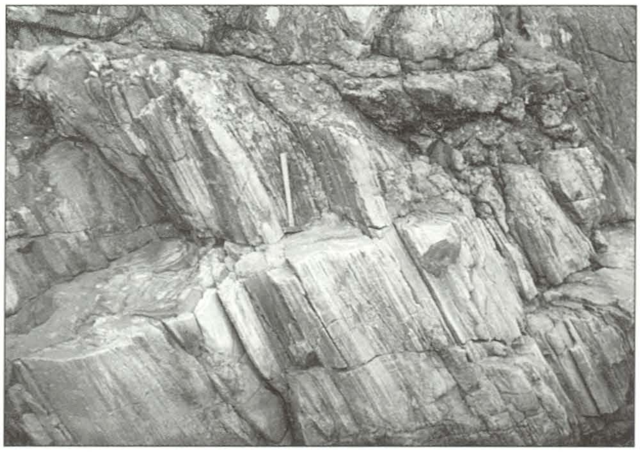

Fig. 14. South-plunging mullion-like L-tectonite in the fold core (loc. 180-530) of the Åpta antiform (F_4), looking east.

display a parallelism of foliation and lithological layering. The only exception is in the fold hinges of the earliest folds.

Lineation, defined as the subparallel to parallel alignment of elongate, linear fabric elements, is common in most of the mafic rocks but rarely observed in felsic rocks except in extremely deformed zones such as fold-hinges. There are several types of lineation, but the two most common are *mineral lineations* and *crenulation lineations*.

The most common mineral lineation occurs when prismatic minerals like amphiboles have undergone a preferred directional recrystallisation along the direction of least strain, normally parallel to the local fold axis. This is the common, penetrative type of lineation in amphibolites and mafic layers in banded gneisses.

Another form of mineral lineation comprises aligned aggregates of mafic minerals such as orthopyroxene, clinopyroxene, hornblende and biotite, which are common in granulite -facies rocks. In rocks with abundant biotite, small elongated and tightly spaced fold hinges, consisting mostly of biotite, are found in an array of relatively straight troughs and crests, always making a measurable lineation. This small-scale *crenulation* (Fig. 12) is the most common type of lineation, parallel to the axes of the related minor folds as well as to the mineral lineation (Fig. 13). The crenulation becomes more accentuated especially in the major fold-hinge zones, where it results in a rod- or mullion-like structure as the rock changes from a S-tectonite to a LS-tectonite, and rarely even to a L-tectonite (Fig. 14).

Geometrical analysis of the major structures

Folding was first recorded in the Precambrian of southern Norway in 1854 by Kjerulf (1879) who recognised folds around Kragerø and Arendal, but he believed that block-faulting was the most important deformation mechanism acting on the gneisses. More recent investigations have shown that the majority of the metamorphic rocks have been deformed by ductile flow under elevated temperature and high confining pressure. This led to passive flow-folds formed during multiple phases of deformation with common refolding of early tight to isoclinal folds, exemplified in the Flekkefjord area and demonstrated on the map (160-550).

The first phase of deformation

The first phase of deformation led to the development of isoclinal to tight folds and the high-grade metamorphic recrystallisation created a penetrative foliation almost always parallel to the lithological layering. The only general exception is the axial-plane foliation in the hinges of the isoclinal folds of the first phase of deformation. These isoclinal folds are only found on a mesoscopic scale as rootless intrafolial folds (Fig. 15). Furthermore, these folds are only found within the banded gneisses and originated during the first (F_1), major, foliation-forming metamorphic event (Table 1).

After this event, the banded gneisses were intruded by the big-feldspar amphibolites and the massive orthopyroxene amphibolites. The intrusive origin of the massive orthopyroxene amphibolites is clearly demonstrated in the Lande area (225-560) where one of them bifurcates and discordantly intrudes across the foliation and lithological layering of the banded gneisses.

The second phase of deformation

It has never been possible to find evidence for a phase of deformation separating the following three major rock-groups; big-feldspar amphibolites, massive orthopyroxene amphibolites and granitic gneisses. They were all deformed together during the first phase of large-scale deformation (F_2) which folded the granitic gneisses, amphibolites and older banded gneisses into large isoclinal folds. The closure of one of these large isoclinal folds is seen south of Åsen (160-550) where the banded gneiss wraps concordantly around the fold-hinge of a south-plunging synform with a steep axial surface (subarea 2, Fig. 16). Within the layers in the enclosing banded gneiss, several small intrafolial isoclinal folds with axial-plane foliation parallel to the general foliation and lithological layering are found all the way around the fold

Fig. 15. Isoclinal intrafolial fold (F_1) with axial plane foliation; looking towards north in the banded gneiss formation along the northern shore of Selura (loc. 110-665).

Deformation phase	Fold style Lineation	Fold axis, orientation	Axial plane	Metamorphism	Age (Ma)
Intrusion of post kinematic granites, charnockites, mangerites, anorthosites and granite pegmatites				980-850	
F6	Open to Gentle	E-W	E-W steep	None	Post 990
F5	Tight to Close	N-S to NE-SW	N to NE steep	Amphibolite facies	990-980
Intrusion of the Homme granite.					990
F4	Tight to Isoclinal	N-S	N-S moderately E-dip	Granulite to Amphibolite facies (W to E)	1050-990
Intrusion of the Feda porphyric granite.				1050	
F3	Isoclinal Recumbent? At present reclined	Variable Originally N-S ?	At present approx. N-S Originally horizontal ?	High-grade	1160-1050
F2	Isoclinal Recumbent?	Variable Originally N-S ?	At present variable Originally horizontal ?	High-grade	1160-1050
Intrusion of gabbroic and granitic plutons (pre 1160 Ma); and the U-Pb dated Hidderskog charnockite					1160
F1	Mesoscopic isoclinal folds	Variable	Parallel foliation	High-grade	Pre-1160

Table 1. Structural and metamorphic evolution and magmatic, intrusive events, in the Flekkefjord area, during the Mesoproterozoic to Neoproterozoic period.

hinge. This suggests that the mesoscopic, F_1 isoclinal folds are older than the macroscopic Åsen isoclinal fold (F_2).

The third phase of deformation

If the trace of the axial surface of the Åsen fold is followed to the north, it runs within the banded gneiss layer which is refolded around the Hafknuten hinge zone (145-705). The F_3 fold (Hafknuten fold) is a large, complex, asymmetric isoclinal fold of reclined type, as the E-W trending fold axis plunges approximately 45° eastwards within the N-S striking and E-dipping (45°) axial surface (subarea 21, Figs. 16 & 17).

The N-S striking and east-dipping foliation in the eastern limb is folded around the hinge-zone, where it gradually turns to an E-W strike and vertical dip when crossing the axial surface. Farther west the strike turns southwards with an easterly dip as the limbs become parallel. The curving axial surface of the Hafknuten fold strikes N-S in the fold-hinge zone, turning gradually to a NE-SW strike towards the south, with a dip of roughly 45° to the east and southeast. The amplitude of the fold is 10 to 15 km.

Towards northwest there is another isoclinal fold (Mjåvatn fold) of the same style and magnitude, although the orientation is rotated about 20° anticlockwise in relation to the Hafknuten fold (subarea 20 and 29, Figs. 16 & 17). In the hinge-zone of the innermost banded gneiss layer of this fold, the foliation is intensely deformed into small-scale isoclinal folds with axial surfaces parallel to the general axial surface of the macroscopic fold. This trend is at right angles to the general trend of the lithological layering, but the gneiss is nevertheless folded around the hinge-closure even though practically all foliation measurements are parallel to the axial surface. These mesoscopic isoclinal folds may be repeated half a dozen times and this repetition of folds is a typical feature in all the banded gneiss units; it is normally impossible to establish whether these folds belong to F_2 or F_3 (Fig. 5).

No intrusive events have been noted between F_2 and F_3. Taking into account the structural style, orientation, magnitude and continuity of the structures, it is reasonable to conclude that F_2 and F_3 formed under similar tectonometamorphic conditions in a continuous major deformational event.

The fourth phase of deformation

The Feda porphyritic granite intruded the metamorphic banded gneiss/granitic gneiss terrain discordantly after F_3 and was subsequently deformed by F_4, recrystallising under amphibolite-facies conditions to an augen gneiss. F_4 also produced tight to isoclinal folds, refolding the older isoclinal folds into the typical pattern of superposed folding described by Ramsay (1958). The original shape of the Feda pluton may well have been quite elongated as it can be followed northwards for more than 100 kilometres. On the other hand, the deformation that folded it into large isoclinal folds, and refolded the older complex into extremely complicated triple fold structures, must have changed the original shape considerably.

The present level of exposure in the southernmost part of the Feda isoclinal fold with the Feda augen gneiss in the fold-core, shows a south-plunging antiform with a steepening of the fold axis to the south (subarea 1 & 5, Figs. 16 & 17). The axial surface strikes N-S with a moderate dip to the

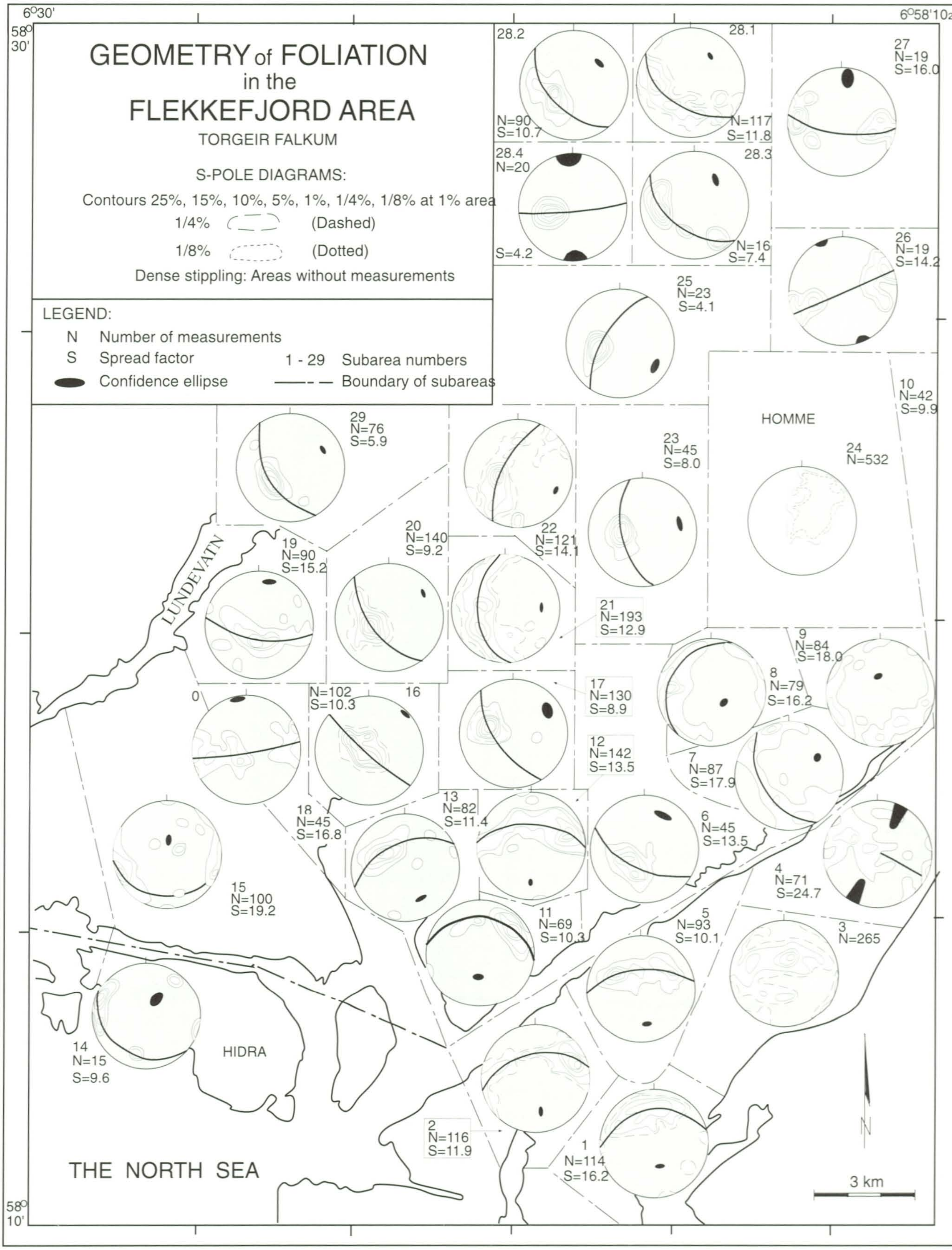

Fig. 16. Equal-area diagrams (lower hemisphere) for the foliation subdivided into 32 subareas.

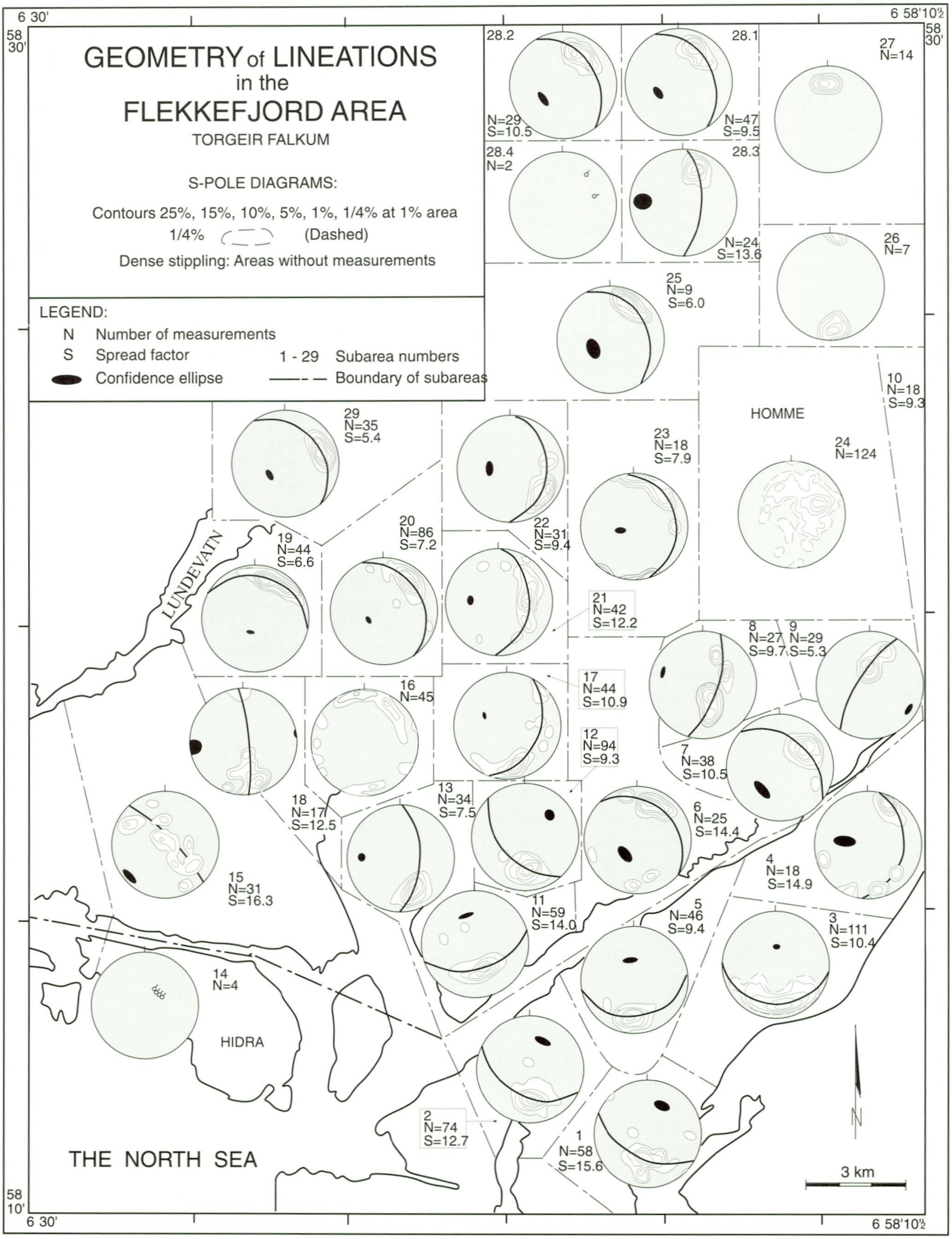

Fig. 17. Equal-area diagrams of linear structures, subdivided into 32 subareas.

east and the limbs become completely parallel a couple of kilometres north of Feda (180-640). Further northwards it splits into two limbs accentuated by the folding; the split itself may represent an original apophysis from the main intrusive body.

The Feda porphyritic granite was strongly deformed and preferred directional recrystallisation developed a strong mineral lineation together with a pronounced crenulation parallel to the main fold axis. The rectangular alkali feldspar megacrysts were reorientated into parallelism with the foliation, and dissolution in the high-strain direction coupled with mineral growth in the low-strain direction gave the typical ovoid or augen-shaped megacrysts (Fig. 11). This is most pronounced in the fold hinge and along the borders, whereas the central part of the body commonly shows almost rectangular and only partly oriented megacrysts. The metamorphic grade was upper amphibolite facies merging into granulite facies immediately to the north of the map area (Hermans et al. 1975, Tobi et al. 1985, Bingen 1989, Bingen et al. 1990, 1993, 1996).

The earlier macroscopic double folds were deformed during F_4 into a triple-folded superposed fold pattern. The westernmost area may serve as an example. To the west of the Hafknuten-Mjåvatn folds, the map shows the typical pattern caused by multiple refolding of older isoclinal folds. The granitic gneiss in the westernmost part of the map can be traced from Stolsfjorden (100-600) northwards to Lundevatn (030-680), where it has been folded into an antiform, plunging southwards below the banded gneiss at Loga (090-695). It appears again immediately to the north of Flekkefjord and forms a dome beneath and around the town. Further to the southeast it is again folded, almost isoclinally, below the banded gneiss and appears again after a few hundred metres, and can then be followed down to Stolsfjorden (095-605) where it crosses the fjord, connecting to the starting point. Traced northwards from south of Flekkefjord it crosses lake Selura (120-660) and is probably connected with the granitic gneiss which was folded isoclinally around the inner banded gneiss unit in the Mjvatn fold. This is the most uncertain connection as an inferred discordance somewhere in the Selura lake can also explain the map pattern.

From the antiform at Loga (090-695) the granitic gneiss can be traced around the hinge of the Mjåvatn fold outside the map area, but it bends southwards again and wraps around the eastern limb of the Hafknuten fold, pinched between the banded gneiss to the west and the Feda augen gneiss to the east. From there it can be followed all the way down to Fedafjord where it crosses the fjord and is folded around the Åpta antiform, disappearing into Fedafjord to the south of Leirvika (220-610). It is probably the same unit which appears again along the eastern border of the northermost part of the Feda augen gneiss. In spite of some uncertainties, the map pattern clearly demonstrates the complex result of superposed folding.

The fifth phase of deformation

After F_4 with high-grade metamorphism, several porphyritic granites intruded South Norway; one of them was emplaced in the eastern part of the Flekkefjord area (240-720). The Homme pluton (Falkum 1976a,b) was emplaced forcefully, tilting the country rocks and ultimately cross-cutting them (subarea 24, Figs. 16 &17). This pluton was deformed and metamorphosed in amphibolite facies during F_5, but the ductile deformation did not erase the discordant contact relationship. The Homme granitic gneiss is classified as a late-kinematic pluton, as it was affected by only the last regional deformation and metamorphism.

This deformation folded the older complex into tight to nearly isoclinal folds along N-S to NE-SW trending fold axes with steep axial surfaces. This variation is presumably a result of the orientation of the earlier structures. The refolding increased in intensity towards the north and east and the reason for the excellent preservation of the older fold-structures is related to the size of the folds. A gradual decrease in fold-amplitude from the older to the younger folds was the most important factor in partly preserving large, older folds from being obliterated during refolding by smaller younger folds. An example of one of these F_5 folds is found in the innermost part of Fedafjord (220-615, Fig. 2a), where the Feda augen gneiss and the overlying granitic and banded gneisses were refolded around a NE-plunging antiform (subarea 6 and 7, Figs. 16 & 17).

The sixth phase of deformation

The last ductile deformation took place apparently without any related metamorphic recrystallisation. It refolded the ol-

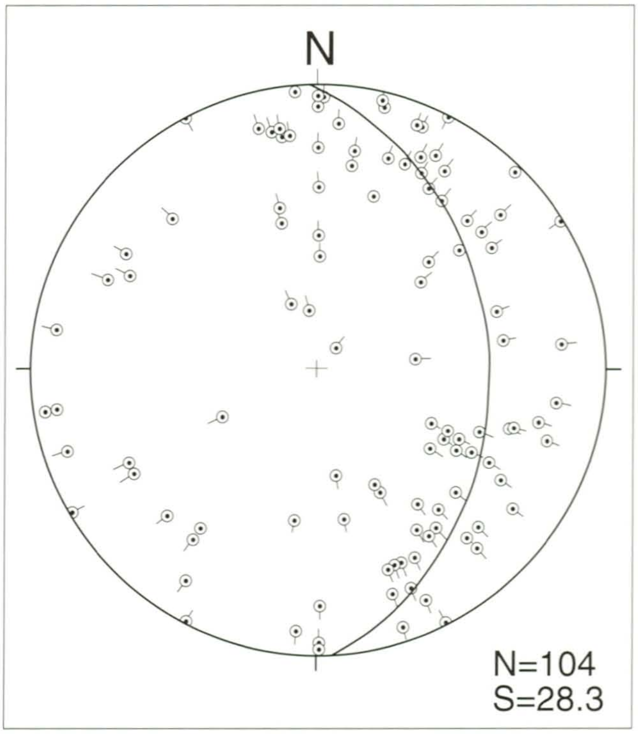

Fig. 18. Equal-area diagram of 104 fold axes from mesoscopic open and gentle folds from the Flekkefjod area. The large spread is demonstrated by the high spread factor (S) of 28.3 when a common π-circle is calculated.

der structures into either open or more commonly gentle folds with steep axial surfaces and near-horizontal fold axes on both mesoscopic and macroscopic scales. The general trend of several of these large-scale undulations is E-W, suggesting a N-S compressional regime with considerably less deformation than in the previous phases of deformation.

On the mesoscopic scale, many open to gentle folds are found with a great variation in the trend of their fold axes (Fig. 18). There is, however, a certain concentration around NE-SW and NW-SE trends. The first trend is parallel to the elongation of the northernmost protrusion of the post-kinematic Lyngdal hornblende granite in the southeastern part of the Flekkefjord area, while the latter trend is parallel to the main axis of the anorthosite-charnockite massifs. Locally, many of these post-kinematic intrusions affected the neighbouring country rocks which became contact metamorphosed and/or deformed or brecciated into agmatites. The foliation around the post-kinematic plutons is normally transposed into a steep attitude, in places up to several hundred metres away from the contacts (subareas 1-4 & 14-15). These observations make it reasonable to suggest that at least some of the open to gentle folds in the neighbouring subareas to the post-kinematic intrusions formed during their emplacement, although no direct link can be established in the field.

If this is correct, some of the younger folds related to this last phase of deformation may represent a posthumous phase in relation to the orogenic movements, as they were formed during the emplacement phase of the post-kinematic intrusions after the horizontal compressive movements had come to a halt. There is, however, a regional set of open folds, commonly with E-W axes and steep axial surfaces, which are older than the supposed posthumous folds and they probably formed during the latest stages of the regional deformation in an apparent N-S compressive regime. These movements represented the waning stage in the collision tectonics of the Sveconorwegian orogeny.

Geometry of foliation and lineations

The structural evolution in the Flekkefjord area can be explained by a combination of at least six separate phases of ductile deformation with plutonic intrusions between some of them (Table 1). A very complicated pattern for the foliations (s-surfaces) and lineations is therefore to be expected. With regard to s-surface orientation, this can be demonstrated (Fig. 16) with the display of all foliation measurements (N=3182) and π-axes for all subareas (Fig. 19).

Lineations also show a considerable spread within each subarea. The area was originally divided into 32 subareas (Figs. 16 &17), which are further subdivided to give a total of 47 subareas. The Homme subarea 24 (Fig. 20) has been further divided into 14 subareas (Falkum 1976a).

A computer program has constructed the great circle girdle of best fit for the foliations within each subarea and calculated the fold axis (π-axis). This is given as a <u>*confidence ellipse*</u> constructed from the standard deviation (SD) of the measurements. Furthermore, the SD of the deviation of points from the great circle is calculated and given as the <u>*spread factor*</u> (Kalsbeek 1966), denoted S in the diagrams.

Comparing Figs. 16 and 17, it is evident that the π-axes and the measured lineations grossly overlap in almost all subareas and discrepancies are readily explained. For instance, the poor fit in subarea 28.4 is a result of a point maximum, and if this subarea is combined with one of the adjoining areas, this misfit would have been avoided.

In a few other instances, the subarea boundaries have been selected in such a way that several phases of deformation interfere, readily seen in subareas 3 and 4, where F_4 is overprinted by F_5. In such cases, the subareas must be further subdivided. The reason for this not being done in Fig. 16 is that the area was subdivided on the basis of the major structures and these subdivisions have been kept undivided in order to avoid manipulation with them after the data had been computed. However, in the further computer calculations, subareas 3 and 4 have been subdivided into three and two new subareas, respectively (3.2, 3.3, 3.4, 4.1 and 4.2 in Fig. 21).

In order to view all measurements in one diagram, the π-axes for all 47 subareas are displayed in Fig. 19 where the confidence ellipses of the π-axes show considerable deviation from the calculated great circle girdle; the deviation is quantified by the high spread factor of 22.2. The orientation

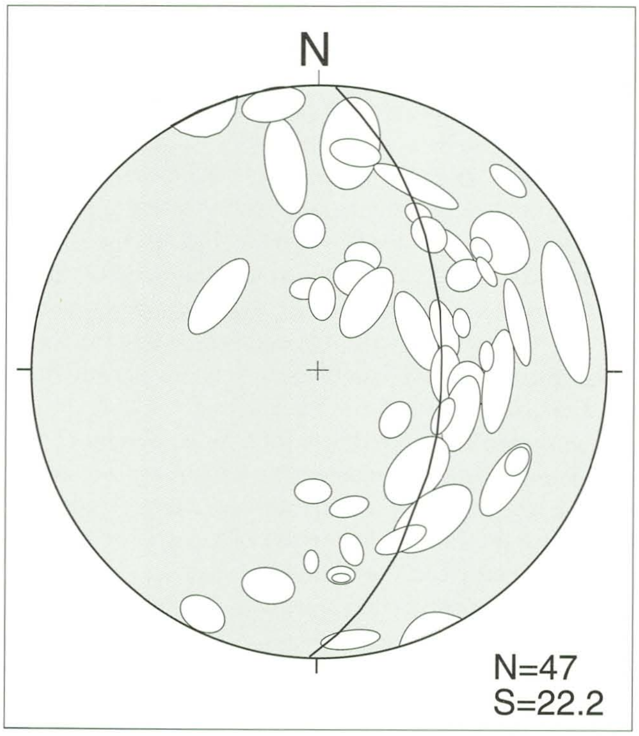

Fig. 19. Equal-area diagram of π-axes from all 47 subareas in the Flekkefjord area, shown as confidence ellipses. There is a large deviation of many of the π-axes from the computed great circle, demonstrated by the spread factor (S) of 22.2.

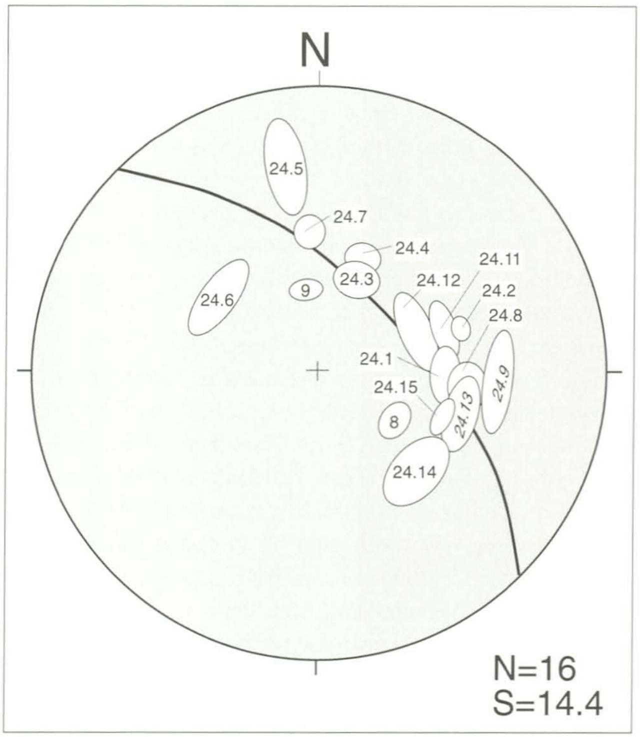

Fig. 20. Equal-area diagram of π-axes from the Homme subarea (no. 24) and the two adjacent subareas to the south (nos. 8 and 9). The π-axes are gradually tilted towards the northwest along the great circle, going from the subareas in the southeast to the subareas in the northwest.

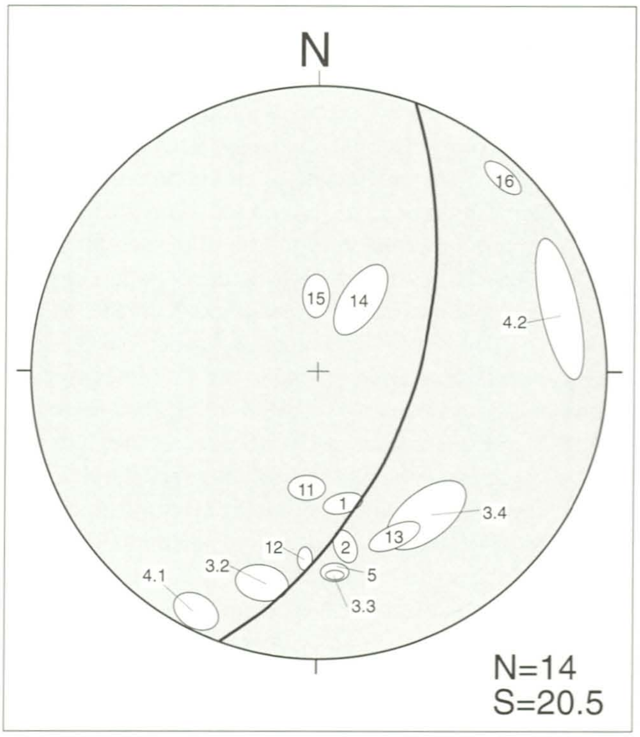

Fig. 21. Equal-area diagram of π-axes from the subareas close to the anorthosite massifs (nos. 14 and 15) and the post-kinematic plutons (nos. 1, 2, 3.2, 3.3, 3.4, 4.1, 4.2, 5, 11, 12 and 13). Subarea 16 represents the Flekkefjord dome.

of the great circle shows that several of these confidence ellipses are situated within a N-S striking and east-dipping plane. There is, however, a certain concentration of confidence ellipses which depart significantly from that plane and these are notably found within three areas, either with SW- or NW-plunge, or with a relatively flat plunge to the ENE (Fig. 19).

It is quite evident from the field investigation that the forceful intrusion of the Homme gneissic granite tilted the country rocks to a considerable extent. This area is covered by subarea 8, 9 and 24; the last being subdivided into 14 new subareas (24.1-24.15), and π-axes from these 16 subareas are shown in Fig. 20. The π-axes from the subareas within the pluton (24.11-14.15) and to the west (24.1-24.2 and 24.8-24.9) are close to the N-S running girdle on the previous diagram (Fig. 19), while the other π-axes (24.3-24.7 and 9) plunge to the N and NW. This is interpreted as a tilt of the country-rock s-surfaces containing E-plunging lineations. They were gradually tilted from E- to NW-plunge in connection with a mushroom-shaped rise of the diapiric magma during its ascent and the result is a total rotation of more than 90° for the most tilted areas (subareas 24.5 and 24.6). The common plane for all these π-axes is steeply dipping to the NE with a NW-SE strike, indicating the trend of the axis of rotation (Fig. 20). Because this movement was caused by local magmatic diapirism, these 16 subareas can consequently be removed from the diagram in Fig. 19.

Other areas which were distorted by intrusive activity are found to the south and southwest on the map. The Lyngdal hornblende granite, the Farsund charnockite, and the Hidra and Åna-Sira anorthosite massifs all influenced their country rocks during emplacement. The π-axes from these areas (Fig. 21) are situated in a NNE-SSW-striking plane steeply dipping to the ESE. The relatively large variation is shown by a high spread factor of 20.5. The almost flat confidence ellipse of subarea 16 represents the culmination of the dome structure around the town of Flekkefjord. The other low-plunging confidence ellipse of subarea 4.2 belongs to a subarea close to the post-kinematic Lyngdal hornblende granite in the southeastern part of the map area. Consequently, all these π-axes have also been removed from the diagram (Fig. 19).

The 17 π-axes from the remaining subareas are plotted in Fig. 22, and the great circle with π-axis was computed. The result shows that the confidence ellipses fit very well to an almost N-S striking plane dipping moderately to the east. The excellent fit is demonstrated by the low spread factor of 8.8.

Subarea 18 (Fig. 22) is only 3 km away from the anorthosite massifs and the extreme western part of this subarea could have been slightly influenced during their post-kinematic emplacement. The southernmost part of subarea 26 and subarea 10 to the east, could also have been slightly tilted during emplacement of the Homme granite, but as no direct observations in the field have justified this assumption, these subareas are plotted together with the other subareas which are considered not to have been influenced by

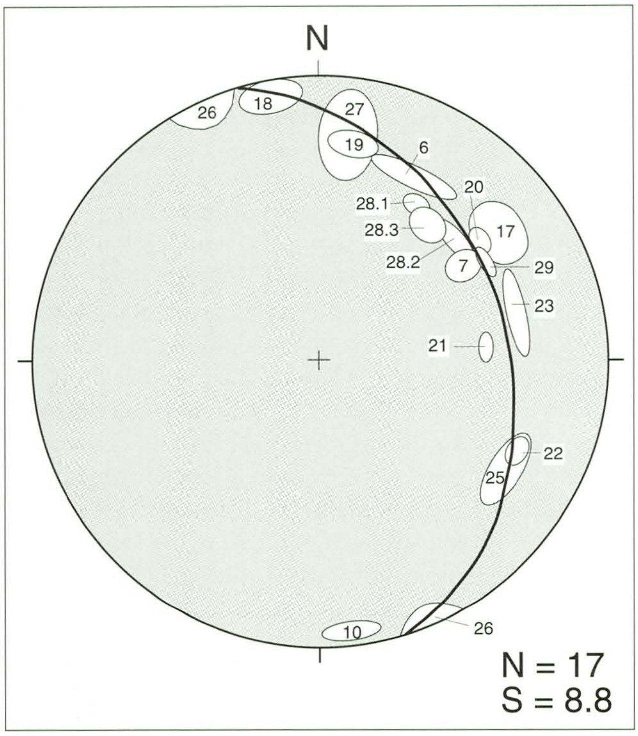

Fig. 22. Equal-area diagram with confidence ellipses from the remaining 17 subareas where the influence of the post-kinematic plutons is negligible or absent . The low spread factor (S = 8.8) of the calculated great circle shows that all π-axes are closely confined to a N-S striking and moderately E-dipping plane.

distortion during emplacement of the late- and post-kinematic plutons.

The interpretation based on this geometric analysis leads to the conclusion that the multiple phases of deformation refolded the older structures roughly around the same fold axes during the development of the macroscopic, isoclinal fold-structures and that the fold axes remained in the N-S trend as long as it was possible to deform the layers into this main trend. Apart from the faint, last phase of deformation, the overall stress field was rather stable during the deformational phases and dominated by an E-W compressive regime. Deviations from this trend occur where the country rocks were tilted by the late- and postkinematic plutons.

The general east-dipping foliation becomes steeper to the east of the map area, ultimately showing a westerly dip which may be interpreted as a large-scale compressive fan structure supporting the observation of an east-west compressive regime.

Kinematic interpretation of the structural evolution

Since the mesoscopic, intrafolial, isoclinal F_1 folds may represent several phases of deformation and recrystallisation, establishing the orientation of the stress field under which they were formed is conjectural. This is also to some extent the case for the F_2 folds, but the reclined folds of the next generation may be unfolded around the F_4 folds. As the F_4 folds are isoclinal, the unfolding will rotate the limbs 90° and the axial surfaces of the F_3 folds are brought to an E-W striking position with the fold axes in a N-S trend. As these folds (F_3) are also isoclinal, the unfolding of the F_2 folds may be 180°, leaving also them with N-S fold axes.

Where the F_4 axes are relatively undisturbed by later folding, they also trend N-S. The trend of the subsequent phase of deformation coincides with this direction apart from in several places where the previous fold-structures apparently prevent it.

The original orientation of the axial surfaces of the macroscopic folds of the first two phases of deformation is interpreted as relatively flat-lying with fold axes dominantly in a N-S trend, suggesting that these folds formed as recumbent folds in an E-W compressive regime. The successive phases produced gradually more upright folds, refolding the older folds to reclined-type folds with east-dipping axial surfaces in the same E-W compressive regime.

Another general trend of development is seen in the fold-amplitudes which become smaller and smaller with time. The kinematic analysis of the fold-structures shows that they successively developed along the same N-S trending fold axis, suggesting an E-W compressive regime which seemingly became stronger with time. On the other hand, the smaller amplitude may be the result of there being less and less space for the folds to develop. This may well have been caused by the piling up of structures in the later stages of continent-continent collision.

Discussion

The structural analysis revealed several phases of deformation which proceeded under high-grade metamorphic conditions, separated by plutonic intrusions, but has not given any clue to the origin of the oldest rocks - the felsic and mafic layers in the oldest part of the banded gneiss. The pronounced layering could possibly be a primary feature, but it is more likely that it represents a secondary structure. The presumed metasediments do not give any clue as to what types of primary bedding might have been present. Considering the strong deformation of superposed folding under extremely ductile conditions, it is likely that bedding transposition has eradicated all traces of primary layering. Several of the younger layers probably intruded as sills and possibly even dykes, subsequently deformed and transposed into complete parallelism with the prevailing layering, so little if anything of the original structures may be found.

If the felsic layers are the oldest, they point to a continental, granitic basement. On the other hand, if the mafic layers are the oldest, they could be relics of basaltic extrusions. If the original rock sequence comprised ocean floor basalts, they were probably obducted onto a continental crust and extensively intruded by granitic magmas, representing a basement for the subsequent formations. This is not the same kind of obduction as proposed by Michot (1984) and Demaiffe & Michot (1985), whereby anorthosite massifs embedded in one continental plate collided with another continental plate and were obducted or thrust onto it.

In the Flekkefjord area, *garnet-cordierite-sillimanite-biotite schists* grading into garnetiferous gneisses and the *quartzite-layered amphibolite* are the only possible metasediments recorded. Both to the west and particularly to the east in the Kristiansand and Bamble areas, thick quartzites, garnet-sillimanite-cordierite-biotite schists and garnetiferous gneisses together with thin marble/skarn horizons are common (Falkum 1966b), indicating an epicratonic shelf type of shallow-water sedimentation. One possibility is that these metasediments also existed in the Flekkefjord area, but were removed by the massive invasion of intrusions combined with the pervasive deformation. Another possibility is that carbonates and quartz-dominated sediments were never deposited in this environment. The garnet-sillimanite-cordierite-biotite schists and garnetiferous gneisses (metapelites and sandy argillites) may then be an environmental indicator, suggesting a clay-dominated sedimentary domain. A deep basin on a continental shelf, like the North Sea graben, or an ocean floor environment are possible candidates and leave the nature of the earliest basement rocks unresolved.

Another question is whether the apparent metasediments were true sediments or were formed by other processes. In some areas, the garnet-cordierite-sillimanite mineral assemblage or similar alumina-rich rocks have been interpreted to represent a residual after partial melting (Leake & Skirrow 1960, Grant 1968, Clifford et al. 1975). One important chemical aspect of the garnet-sillimanite-cordierite-biotite schists and garnetiferous gneisses in the Flekkefjord area are the high K and Al contents and the very low content of Ca. The latter would not be the first element to be removed if the original rock had undergone partial melting. Nesbitt (1980) calculated the chemical result after removal of 10, 20 or 30% of granitic melt from an average amphibolite-facies rock and showed that the amount of Ca in the residual rock increased as more liquid was formed and removed. Furthermore, Ca is not a normal major constituent in clays, so the chemical composition of the Ca-poor Flekkefjord rocks strongly points to a clay protolith.

The origin of the *quartzite-layered amphibolite* rock is more difficult to interpret since several phases of strong deformation and high-grade metamorphism have transformed all primary structures in the protolith. It could represent quartz-sand alternating with basaltic ash layers. Another possibility is that this peculiar rock formed by metamorphic segregation during strong tectonism and high-grade metamorphism, or that the quartz layers simply represent several, cm-thick, quartz veins deformed into conformity with the foliation and profoundly altered by complete recrystallisation.

Thus, relatively early during the evolution of the Flekkefjord rock complex, a banded gneiss suite containing metasediments was formed and subsequently folded and metamorphosed under high-grade conditions during the first phase of recognisable deformation. This early banded gneiss was subsequently intruded by gabbroic magmas, described as the *big-feldspar amphibolites* and the *massive orthopyroxene amphibolites*. One of the latter amphibolites clearly cross-cuts the F_1 foliation in the Lande area, but was later folded during the F_2/F_3-phases of deformation.

The *big-feldspar amphibolites* also intruded the banded gneiss complex at a very early stage, and were subsequently deformed and metamorphosed during several phases of deformation. The individual lenses are always concordant with the surrounding banded gneiss. Realising, however, how difficult it is to separate the different phases of deformation on a mesoscopic scale, the emplacement may have taken place after the first foliation-forming event, but before the second or third phase of large-scale folding. According to this interpretation, their emplacement occurred in the same period as that of the massive orthopyroxene amphibolites and before the granitic gneisses.

Another interesting feature related to the big-feldspar amphibolites is that they signal the presence of anorthosite kindred rocks somewhere at depth several hundred million years before the emplacement of the large anorthosite massifs. The big-feldspar amphibolites, and in particular the massive orthopyroxene amphibolites, are probably high-level expressions of giant mafic magma chambers accumulated in the Moho region. Such enormous amounts of magma underplating the crust would obviously have been an important heat source. Vertical movements of magma would inevitably lead to convective heat transfer, which is far more efficient than conductive heat transfer, facilitating large-scale partial melting of overlying crustal rocks. This could explain the enormous amounts of granitic magmas emplaced as batholiths, plutons and also injected as sills into the banded gneiss complex and later deformed and recrystallised to gneisses. Between 200 and 250 Ma later, during the last phase of the Sveconorwegian orogeny, enormous magma reservoirs again underplated the crust, leading to the emplacement of the Egersund anorthosite-mangerite-charnockite complex in southwest Rogaland to the west of the Flekkefjord area. It is also probable that underplated magma reservoirs played an important role in connection with the formation of all the post-kinematic plutons emplaced in the Flekkefjord area and elsewhere in southern Norway.

Interpretation of the emplacement of the remaining rock-types in the Flekkefjord area seems fairly straightforward. They all formed as the result of magmatic intrusions and their variation is a result of different compositions and PT-conditions during ascent, emplacement and crystallisation. Their post-crystallisation history was also important in some cases. All, except the post-kinematic rocks, were profoundly altered by dynamic recrystallisation during the metamorphic episodes.

The *kinematic evolution* gave rise to large-scale passive flow folds, refolded several times during multiple phases of deformation. Subsequent refolding occurred along the same N-S trending fold axes. The early large-scale isoclinal folds with amplitudes of several tens of kilometres, were most probably formed as recumbent folds with flat-lying axial surfaces and N-S trending fold axes. Later refolding rotated them around N-S axes, which tilted them to reclined folds with N-S striking axial surfaces combined with E-plunging fold axes and a parallel E-dipping foliation. Such large-scale isoclinal folds are known from many mountain chains, formed during horizontal transport in connection with compressive orogenic deformation and/or gravity-induced vertical movements, where the folded strata were tilted to a horizontal position during the later stages (Talbot 1974).

All folds, apart from those formed during the last phase of deformation, developed roughly co-axially and with gradually smaller amplitudes. The later folds were formed on the semi-planar limbs of earlier folds and the refolding of older folds commonly resulted in a complicated outcrop pattern, first documented in detail by Ramsay (1958) and later becoming a standard in the literature (Turner & Weiss 1963, Whitten 1966, Ramsay 1967, Ramsay & Huber 1987, Davis & Reynolds 1996).

Refolded folds commonly form in the deeper part of the orogenic core zones in rocks with a general high mean ductility and low ductility contrast, reworked during prolonged orogenic activity. The model of Den Tex (1963) where the deeper parts of orogenic zones have longer periods of deformation than the higher parts, is applicable to the Flekkefjord area. Elevated temperature and high confining pressure in the orogenic core zone normally led to passive folding, as described by Donath & Parker (1964).

Carey (1953) introduced the concept of *rheid* for extreme flow in rocks under small values of differential stress and this concept is applicable to understand the behaviour of the Flekkefjord rocks during at least the first four phases of deformation. Another factor contributing to facilitate passive flow is local partial melting which certainly occurred on a small scale throughout the area. Deformation enhanced by melt-lubrication in connection with batholith emplacement has been described by Andersen et al. (1991).

Multiple phases of deformation are presently known from many regions in southern Norway, although it may be difficult to demonstrate successive refolding on the map-scale. One exception is the work of Huijsmans et al. (1981) from an area just a dozen kilometres to the northwest of the Flekkefjord area. They demonstrated four phases of deformation which all led to the development of mappable macroscopic folds. The first phase (their D_1) produced a macroscopic isoclinal fold-closure with axial surface foliation. D_1 is evidently equivalent to F_1, although only mesoscopic folds of this generation were found in the Flekkefjord area.

The D_2 phase of Huijsmans et al. (1981) can be correlated with the F_{2-3} phases. Large-scale refolding of isoclinal folds has not been recorded to the north of the present area, possibly because isoclinal refolding of earlier isoclinal folds will obliterate all older fold-hinges. Huijsmans et al. (1981) described the next generation of folds (D_3) as open to tight structures which apparently correlate with F_4, leaving F_5 in the Flekkefjord area without any correlative to the northwest. As previously described, this phase of deformation decreases in intensity towards the west and may be completely absent northwest of Flekkefjord.

There is general agreement that the last phase of deformation produced gentle to open folds along E-W axes with amplitudes up to several kilometres (Falkum 1966a, Huijsmans et al. 1981). The excellent correlation of the structures in these two areas situated between the Rogaland anorthosite province and the Agder migmatite terrane suggests that the general kinematic evolution in this transitional zone progressed in a comparable way over a large region. Farther southwards the Lyngdal hornblende granite and the Farsund charnockite were directly emplaced into this zone. Detailed structural correlation with the area to the southeast has not been possible, although refolding of large-scale isoclinal folds around N-S trending axes (Petersen 1973, 1977) is similar to the F_{3-4} evolution in the Flekkefjord area.

This transitional zone between the anorthosite massifs to the west and the Agder migmatite terrane to the east is considered to be an important part of the orogenic accretion zone, defined as the core zone (Falkum & Petersen 1980, Falkum 1985) where enormous amounts of magma intruded an older complex. The origin and evolution of these magmas are believed to be related to a N-S oriented subduction zone situated somewhere to the west of the present coast of Southwest Norway and associated with an E-W compressional regime.

The *metamorphic grade* was granulite facies to upper amphibolite facies during the latest episodes of metamorphic recrystallisation. It is difficult to judge the amount of subsequent recrystallisation of the earlier mineral assemblages. The minerals parallel to the axial planes of the F_1 folds, or parallel to the F_2 and F_3 lineations, all show granulite-facies mineral assemblages in the mafic rocks, probably recrystallised during the pervasive high-grade recrystallisation related to the F_4 phase. This recrystallisation resulted in a granulite-facies mineral assemblage in all mafic rocks within the Flekkefjord area, whereas the felsic or leucogranitic rocks contain hypersthene to the west of a N-S line approximately through the town of Flekkefjord, mapped in more detail farther northwards by Hermans et al. (1975) as the hypersthene line. This metamorphism (during F_4) presumedly recrystallised all previous mineral assemblages, which were probably already at high grade.

The effect of the last deformation (F_5) and accompanying amphibolite-facies metamorphism seems to vanish westwards, whereby the previous mineral assemblage (F_4) survived to a large extent. The mafic rocks in the eastern part of the Flekkefjord area have obviously partly recrystallised under amphibolite-facies conditions, commonly seen as a strong alteration of hypersthene and extensive late crystallisation of biotite.

Temperature estimates for the M2 metamorphism from the area to the north range from 700 to 1050° C close to the anorthosite massifs (Jansen et al. 1985, Tobi et al. 1985). If this estimate is extrapolated southwards, the temperature range during the metamorphic peak in the Flekkefjord area during the F_4 phase, would have been in the range of 700 to 750° C. At a temperature of around 700° C, the geothermometer/barometer of Currie (1971, 1974) for the garnet-cordierite-sillimanite-quartz assemblage can be used. The Fe/(Fe + Mg) ratio in the Flekkefjord schists and garnetiferous gneisses varies between 0.79 and 0.84 (one sample 0.96), giving a pressure in the range of 4 to 5.5 kb. With lower temperature the pressure could be around 6 kb, while higher temperature would point to a lower pressure, possibly as low as 3.5 kb. Since most of these schists are located in the central part of the area, it is reasonable to relate this assemblage to the F_4 phase of deformation and accompanying metamorphism, although it cannot be excluded that it represents the last regional metamorphism. To the north, Jansen et al. (1985) estimated that the M2 mineral assemblage reequilibrated to the M3 assemblage in the temperature range of 550-700° C and 3-5 kb total pressure, and it is possible that this phase is equivalent to the F_5 phase of deformation and accompanying amphibolite-facies metamorphism, particularly in the eastern part of the Flekkefjord area.

The inferred total *pressure* during the recrystallisation in connection with the F_4 phase of deformation was in the range of 4 to 5.5 kb, approximately correlated to a load pressure equivalent to a depth of 15-20 km. The present thickness of the crust is around 30-35 km (Balling 1990), which indicates that the total thickness was 50 ± 5 km, which is a moderate thickness compared to a highly overthickened crust. The thickness of the crust seems to increase eastwards and northwards (Balling 1990) where it becomes more comparable to the overthickened crust in active orogenic zones.

The *absolute timing* of the older events in the Flekkefjord area is difficult to establish, as there are no radiometric age determinations from these rocks. The oldest rocks to the north have ages between 1.5 and 1.4 Ga (Pastels & Michot 1975, Versteve 1975, Verschure 1985). According to Menuge (1988), the crustal precursors of the high-grade gneisses and probably also the amphibolites, were derived from a depleted mantle source in the period 1.5 - 1.9 Ga ago. Consequently, the old banded gneiss units and the metasediments are probably of pre-Sveconorwegian origin, although an early Sveconorwegian age cannot be completely ruled out.

The big-feldspar amphibolites and the massive orthopyroxene amphibolites together with the granitic gneiss units are older than the Hidderskog charnockitic gneiss (Lien 1970), since this charnockite pluton intruded the banded gneiss and one of the granitic gneiss layers. The emplacement of the Hidderskog charnockite was dated to 1160 Ma (Zhou et al. 1996) and it was emplaced before the F_4 phase of deformation as it is folded along the N-S trend. The western part of the pluton makes up a synform which can be traced into the surrounding rocks. This phase of deformation has deformed an already existing foliation within the Hidderskog pluton; this foliation is interpreted as having formed during the F_2/F_3 phases. Consequently, all these rocks were deformed during the second and third phases and subsequently cut by the Feda augen gneiss, dated to 1050 Ma (Bingen pers. comm. 1996, Bingen & van Breemen 1998), constraining the F_{2-3} phases to the 110 Ma interval between 1160 and 1050 Ma. Farther eastwards a major metamorphic episode seems to have been active between 1100 and 1090 Ma (O'Nions et al. 1969, O'Nions & Baadsgaard 1971, Priem et al. 1973, Pedersen et al. 1978, Wielens et al. 1981, Demaiffe & Michot 1985, Verschure 1985). Precise age datings to constrain this phase in the present area are still required, but it could also be close to 1100 Ma.

The F_4 phase occurred between 1050 Ma and 990 Ma, the younger age being constrained by emplacement of the Homme granite (Falkum & Pedersen 1978). Menuge (1988) indicated a possible disturbance of the Sm-Nd system during granulite-facies metamorphism and migmatisation at 1035 Ma ago, and this could be the effect of the metamorphism accompanying the F_4 phase. The last regional deformation and amphibolite-facies metamorphism is placed in the narrow gap between 990 Ma and the earliest post-kinematic pluton at 980 Ma (the Holum granite; Wilson et al. 1979).

According to the U-Pb and Rb-Sr data, the *post-kinematic intrusions* were emplaced between 980 Ma and 900 Ma, and there seems to be a peak around 950-930 Ma (Wielens et al. 1981, Schärer et al. 1996). Some pegmatites are still younger (Neumann 1960, Broch 1964). The Rb-Sr data give in many cases younger ages than the supposed intrusive ages of the U-Pb data, possibly due to slow cooling rates in the deep-seated environment. The difference for the Farsund charnockite, which is 874 Ma (Rb/Sr) compared with 930 Ma (U/Pb), could be caused by enhanced element migration in the catazonal environment because the system remained open to Rb and Sr migration after closure of the U/Pb system in zircon (Pedersen & Falkum 1975).

In spite of being emplaced after the last regional deformation and metamorphism, all the post-kinematic plutons seem to have recrystallised. Their static or post-tectonic recrystallisation is suggested by a stable mosaic texture of mainly quartz and feldspars characterised by grain boundaries with dihedral angles of 120° and by replacement reactions. The original texture within the anorthosites seems to have been very coarse grained. Within large zones, especially close to the contacts, recrystallisation has transformed this into a fine- to medium-grained granular texture. Considering the catazonal environment into which these plutons were emplaced, it is probable that the temperatures in the country rocks were only a couple of hundred degrees lower than in the crystallising magmas. Generally high temperatures, combined with a medium to slow cooling rate of the rock complex, promoted extensive recrystallisation in all post-kinematic plutons.

These plutons were the last massifs to stabilise the South Norwegian craton, which after uplift and erosion became rigid enough to allow dolerite dykes to intrude the basement in the period between 800 and 630 Ma (Storetvedt & Gidskehaug 1968, Versteve 1975, Sundvoll 1987). Some may have intruded even later, as Late Carboniferous or Early Permian dolerites are found farther east (Halvorsen 1970). The dyke intrusions point to a cratonised basement which had suffered considerable uplift and erosion. The uplift must have been less than 20 km in 100 Ma (0.2 mm/year) and the rate of cooling probably in the order of 3 to 6 °C/Ma, which is a medium to slow cooling rate.

Summary and petrogenetic conclusions

The _banded gneiss suite_ is the oldest unit in the Flekkefjord area, and it contains a mixture of various lithologies with different age relations and origin. The earliest banded gneiss units may either be continental felsic gneisses, pervasively dissected by mafic intrusives, or basalts with later granitic intrusions. The original structures were profoundly altered by bedding transposition during deformation, and infiltration of melts formed by local anatexis and remobilisation and injected along the prevailing layering. Later additions of magmas from external sources, intruding as sills and/or dykes, also added a considerable amount of material to the banded gneiss complex. However, multiple phases of intense deformation and high-grade metamorphism have transformed the original rock-types into typical gneisses, obscuring their primary origin.

Sedimentary clay formations were probably deposited on the oldest part of this banded rock complex and later metamorphosed to _garnet-sillimanite-cordierite-biotite schists_ and _garnetiferous gneisses_. The _quartzite-layered amphibolite_ was either sand deposits mixed with volcanic ash layers or formed by local quartz segregation in response to metamorphic and/or tectonic conditions, or quartz veins transposed into complete conformity with the foliation.

The older part of the banded gneiss complex containing the metasediments was deformed during the F_1 phase of penetrative deformation resulting in rootless intrafolial isoclinal folds with axial plane foliation and metamorphosed under high-grade conditions, probably in granulite facies. The crustal precursors to the high-grade gneisses, according to Sm-Nd model ages (Menuge 1988), were derived from a depleted mantle source in the period 1.9 to 1.5 Ga ago, which suggests that the formation of the mesoscopic intrafolial folds and early metamorphism probably took place later than 1.5 Ga, either during a Mid-Proterozoic orogeny or during an early phase of the Sveconorwegian orogeny.

After this episode, the complex was invaded by _mafic sills_ and _dykes_ of gabbroic composition, recorded as the intrusive _big-feldspar amphibolites_ and the _massive orthopyroxene amphibolites_. The primary magmas to these rock-types are considered to have been formed by partial melting, probably in the mantle, later undergoing differentiation during ascent representing high-level expressions of mafic magma which underplated the crust in the Moho region. This is the initiating process in a long-lasting evolution leading to the intrusion of different plutons and batholiths during the Sveconorwegian orogeny.

After the different types of mafic magmas had intruded the banded gneisses and crystallised as gabbros/norites or dolerites, and before they suffered any kind of deformation and metamorphism, huge amounts of _leucogranitic magmas_ were emplaced and crystallised to _biotite granites_. They formed by partial melting of either the underplated gabbroic rocks or the gabbros/amphibolites in the overlying crust, which must have received an enormous amount of heat from the underlying magmas. Generation of the vast amounts of these mafic and felsic magmas probably took place during a period of extensional tectonism, suggesting decompression partial melting. These rocks were emplaced before 1160 Ma when the Hidderskog charnockite intruded the older complex (Zhou et al. 1995).

After 1160 Ma the tectonic style changed to a compressional regime and the entire rock complex was deformed over a considerable period of time, as large-scale refolding occurred, poorly constrained to the period 1160-1050 Ma, possibly around 1100 Ma. The banded gneiss suite with metasediments and the intrusive granitic batholiths and plutons were deformed together during F_{2+3}, resulting in large-scale _recumbent isoclinal folds_. The style and size of the fold structures reflect large-scale horizontal movements, probably along an E-W trend, and folded around N-S axes. Such structures are typically formed under orogenic conditions, both due to a compressional stress field and also in connection with gravity-driven vertical movements leading to mushroom-shaped recumbent folds (Talbot 1974).

The next plutonic intrusive event (at 1050 Ma) was the emplacement of a large _porphyritic batholith of granitic composition_, subsequently deformed (F_4) by isoclinal folds with N-S-trending fold axes and metamorphosed in the period 1050-990 Ma (1035 Ma?), recrystallising to the _Feda augen gneiss_. The previous recumbent isoclinal folds were refolded to moderately dipping reclined folds, resulting in a complicated triple fold pattern.

All mafic rocks were recrystallised within the hypersthene isograd (granulite facies) as were the leucocratic rocks to the west of Flekkefjord, where a N-S trending hypersthene-in isograd was established (Hermans et al. 1975). The _temperature_ during this high-grade metamorphism was in the range of 700-750° C and the _pressure_ between 4 and 6 kb. According to the classification of Zwart (1967a, b, 1969) this orogenic development represents a _low-pressure type_ with cordierite-sillimanite as the typical mineral assemblage in peraluminous rocks.

These F_4 structures were cut by the Homme granite, emplaced at 990 Ma. This granite was deformed (F_5) and metamorphosed under amphibolite-facies conditions in the period between 990 and 980 Ma. The effect of this F_5 phase seems to be rather limited in the western part of the Flekkefjord area, but increased in intensity eastwards and northwards.

The last phase of deformation (F_6) produced open to gentle folds with steep axial surfaces in an apparently N-S compressional regime. They are commonly large-scale folds with amplitudes of up to several kilometres and generally with E-W axes, probably formed during the waning stages of the regional orogenic movements. There may, however, be more than one phase of these late folds, as some of them may be associated with intrusive activity during the emplacement of the anorthosite batholiths and/or the post-kinematic plutons. The kinematic analysis demonstrates a strong influence from these plutons on the neighbouring country rocks (Figs. 20 & 21).

The post-kinematic Lyngdal hornblende granite intruded (950 Ma) discordantly into the banded and granitic gneiss layers in the southeastern part of the Flekkefjord area. It was subsequently intruded (940-930 Ma ?) by the Farsund charnockite (Falkum et al. 1979), which was possibly itself intruded (930 Ma) by the Hidra leuconorite-anorthosite (Demaiffe & Michot 1985). Alternatively, the Farsund charnockite and the Hidra anorthosite could have been in the magmatic stage more or less simultaneously, together with the large anorthosite massifs and jotunite-mangerite-charnockite suite which also were emplaced around 930 Ma (Schärer et al. 1996).

The Lyngdal hornblende granite probably formed by partial melting of the Ti-rich mafic rocks, possibly amphibolites, since it has three times as much Ti as average granites (Falkum 1972). Many gneissic and amphibolitic inclusions, especially along the borders, suggest assimilation of country rocks, another possible process which could lead to an increased amount of such an element as titanium in the granitic magma, although it can hardly explain the large total amount.

The Farsund charnockite could have been formed by partial melting of crustal material under H_2O-deficient conditions, but with CO_2 present. Noritic residues after the formation of anorthositic rocks may have been the protolith which underwent anatexis yielding a dry, Mg-poor charnockitic magma. Crystal settling in a jotunitic magma may also lead to a mangeritic and ultimately to a charnockitic magma (Duchesne & Wilmart 1997). Numerous more or less assimilated inclusions of country rocks in the charnockite suggest some later crustal contamination.

The Precambrian of southern Norway was cratonised during the Late Proterozoic and transected by dolerite dykes in the period between 800 and 600 Ma, revealing uplift and erosion of this old mountain chain.

The conclusion of Barth (1933) that «......many a chapter of the earliest geologic history of the Earth will be wrested from these heatherclad hills......», is still more valid than ever for this marvellous terrain at the southern tip of Norway.

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