The Early Major Structure and Petrology of Rocks in the Bamble Series, Søndeled-Sandnesfjord, Aust-Agder

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The geology and major fold structures of part of the Bamble Series are described from an area near Risør, Aust-Agder (some 15 km SW of the Levang paragneiss dome). Major structures are shown to have formed prior to the intrusion of Sveconorwegian 'hyperites' and seem to belong to the earlier Svecofennian orogeny. The subsequent Sveconorwegian tectonism modified the older structures and caused deformation around the newly intruded 'hyperites'. The Svecofennian complex consisted largely of supracrustals and granitic gneisses, including those of the Levang paragneiss dome, just south of Kragerø. Immediately southwest of this, major structures are complex and disordered due to the development of basement-cover tectonics during the Sveconorwegian orogeny, which affected an area stabilised locally both by this dome and by some of the larger hyperites.

The Sveconorwegian metamorphism amphibolitised the hyperites, remobilised some granitic rocks (with late-stage pegmatites) and produced isolated ortho-amphibole rocks.

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Introduction

The Bamble Series, first comprehensively described by Bugge (1943) trends NE-SW along Norway's Skagerak coast and is separated from the Telemark Suite inland (to the northwest) by a 'Friction Breccia'. To the north of Risør, the coastal tract has tight major folds on NE-SW axes, becoming tighter towards the 'Friction Breccia' with axial planar shearing producing a planar belt striking NE-SW and dipping steeply SE (Touret 1968; Morton et al. 1970; O'Nions & Baadsgaard, Fig. 1. 1971).

In the coastal area around Søndeled and Risør (Fig. 1), there is a regional change in strike from NNE-SSW to E-W. This is complicated by supracrustal-granitic gneiss domes and basins, later intruded by 'hyperites' thought to represent the early stages of the Sveconorwegian Regeneration. This orogeny amphibolitised the hyperites, locally mobilised granitic material and formed numerous late pegmatites.

South of Kragerø (Fig. 1) lies a large dome of granatic paragneiss (the Levang granite gneiss dome). The area at present being considered lies some 15 km southwest of this structure and represents a large fold closure around it. The intervening area, immediately southwest of the Levang dome, has a complex major structure (Fig. 1). It seems to represent a restricted area of typical basement-cover tectonics, with the dome and later, large hyperites

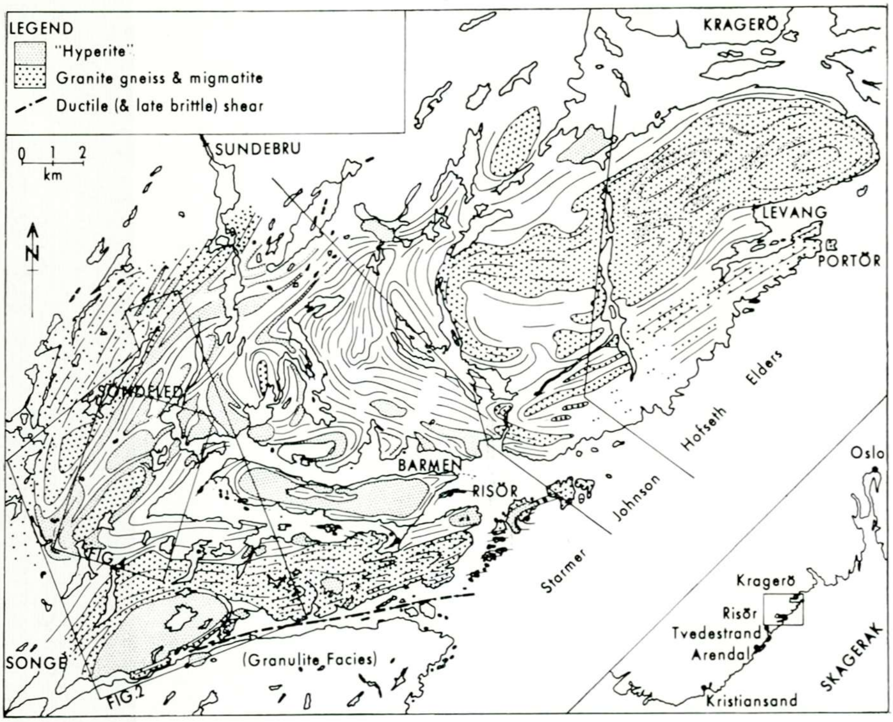

Fig. 1. Locality map, showing general supracrustal structures, main areas of granitic rocks and some hyperites. (Including information from Hofseth 1942, Elders 1964 and Johnson 1976, as shown).

acting as a more rigid, passive 'basement' and supracrustals between behaving as a less viscous 'cover' during the Sveconorwegian orogeny. A considerable amount of ductile shear occurred and was particularly marked in areas not stabilised by large hyperites (Starmer, in preparation). Further southwest, the deformation pattern was more ordered in the area at present under consideration (Figs. 2–5).

South of Risør, the major structure was truncated by a syn-metamorphic ductile shear (Fig. 1) which brought up a dominantly Granulite facies complex to the south (Starmer 1972b). Later detectable movements along this shear occurred after the Sveconorwegian orogeny and involved Greenschist retrogressions. Isolated developments of Hornblende-Granulite facies rocks are found north and west of this shear zone (Touret 1968; Morton et al. 1970; Johnson 1976) but they are not developed on a comparable, regional scale.

The supracrustals

The supracrustals represent the oldest rocks in the area and consist of quartzites, variable biotite-quartz-plagioclase gneisses and schists, sillimanitic and 'nodular' rocks, mixed gneisses and actinolitic gneisses. Thin, concordant amphibolite layers intercalated in parts of the sequence suggest basic metavolcanics within the metasediments; some admixture of material in certain horizons produced biotite-hornblende-quartz-plagioclase gneisses. The individual lithologies form broad bands but with some gradations along and across the strike: rapid alternations of thin layers gave rise to the mixed gneisses whilst thicker units yielded banded gneisses. All types of supracrustal may grade into granitic gneisses and may show varying degrees of granitisation and microline porphyroblastesis.

Quartzites

Quartzites occur as discrete, broad bands and as thin layers in the other metasediments. Many are very pure, but frequently they contain minor amounts of biotite, muscovite, sillimanite, plagioclase (oligoclase), magnetite and more rarely hornblende. Retrogression has formed chlorite, epidote and kaolin and accessories may include zircon, apatite, tourmaline (schorl), ilmenite, rutile and haematite.

The quartzites are massive or weakly foliate with flattening of quartz crystals and alignment of lepidoblastic minor and accessory constituents. Colour banding has sporadically developed from a variable distribution of accessories.

Variable biotite-quartz-plagioclase gneisses and schists

These rocks show all modal variations and may contain almost monomineralic layers of biotite schist. In the present area, quartz-plagioclase rocks (poor in biotite) are extremely rare, except in the mixed gneisses. The plagioclase varies in composition from oligoclase to andesine. Common minor components are muscovite and retrograde chlorite, epidote and sericite. Accessories include apatite, zircon, magnetite, pyrite, sphene, ilmenomagnetite and sometimes sillimanite. Some bands contain major or minor hornblende and show all gradations with thin, concordant amphibolites. Biotite in these rocks is partially retrograde from hornblende. Differentiation has sporadically produced lenses of amphibolite or hornblendite.

Garnetiferous bands are common with small, clear almandines or larger poikiloblasts (Alm_{57-71} Pyr_{5-26}). Some thin layers and lenses contain a little cordierite (usually less than 5 modal %) and this is often accompanied by acicular sillimanite and phlogopitic biotite. Thin graphitic horizons are developed sporadically.

Rocks of this group, particularly the more schistose varieties, were relatively incompetent and were easily squeezed in fold hinges, easily sheared and easily eroded.

Sillimanitic and nodular rocks

Within the supracrustals, biotite–quartz–sillimanite–plagioclase gneisses and schists are extensively developed and some quartzites also contain sillimanite. In some rocks, quartz–sillimanite segregations have formed distinctive 'nodules' which tend to project from weathered surfaces. The 'nodular rocks' may occur

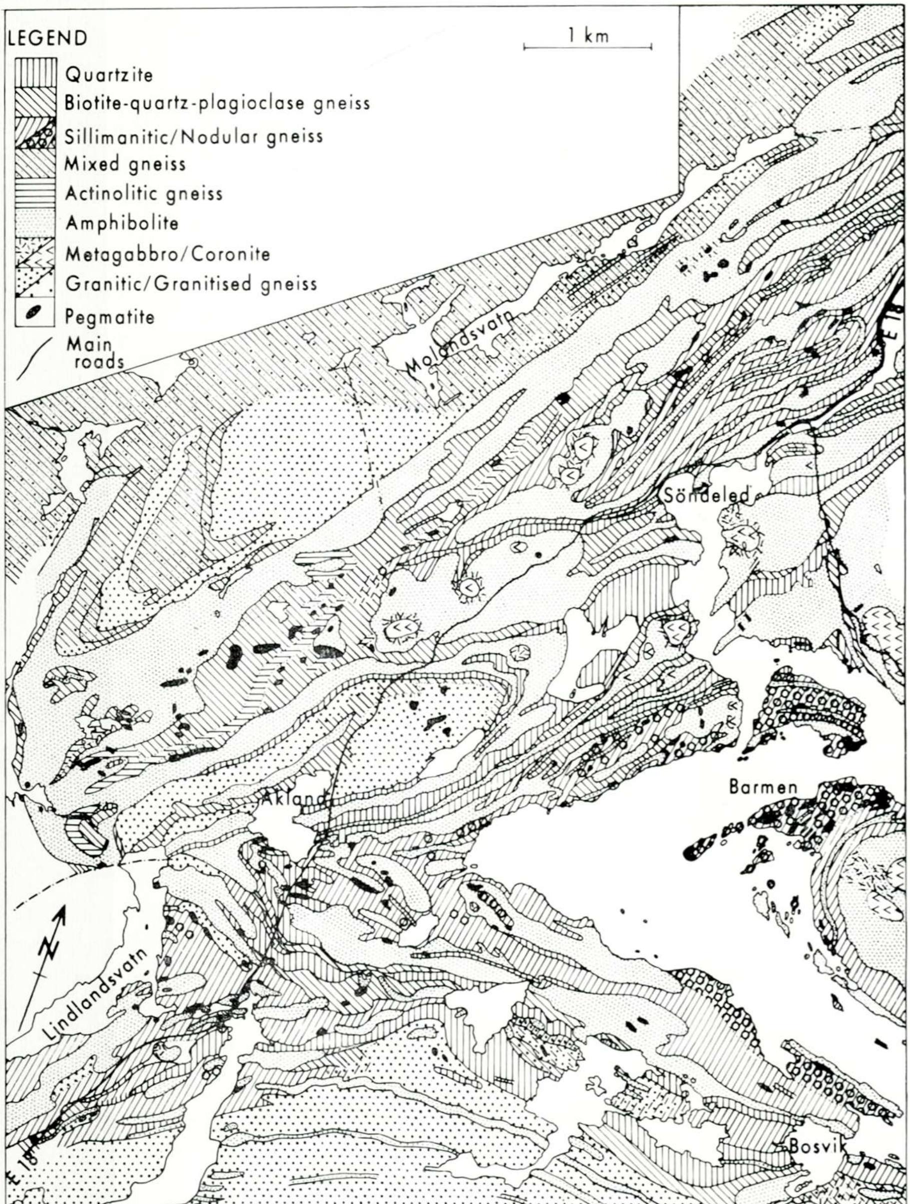

Fig. 2. Lithological map of the area from Søndeled to Laget.

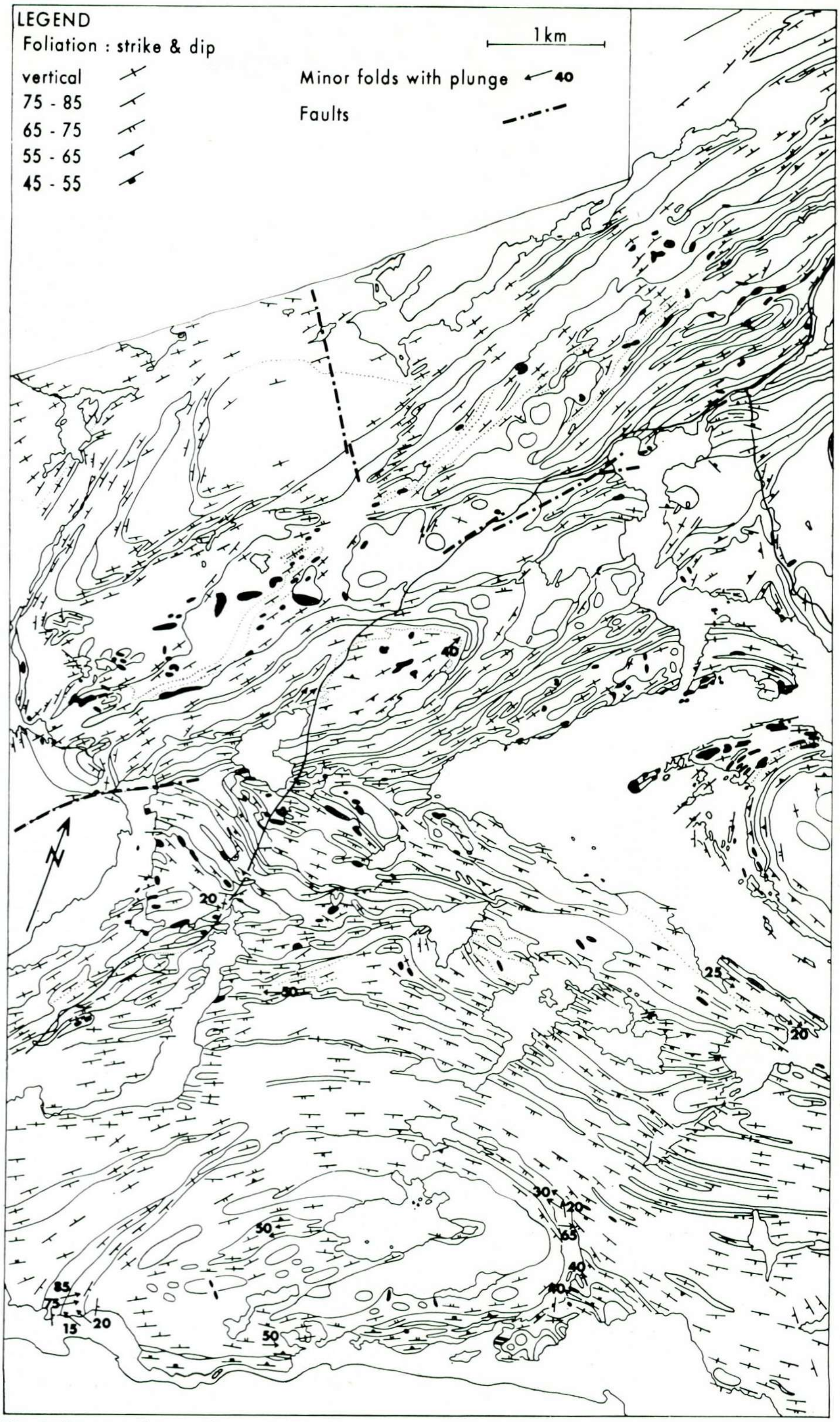

Fig. 3. Structural map of area in Fig. 2.

in isolated patches or as major developments (shown on Figs. 2 and 4). The distribution and field relationships indicate that all types of sillimanitic rock are part of the metasedimentary sequence.

The sillimanite occurs as acicular crystals (up to 1.5 cm in length) and as felted fibrolite masses, the latter often being dominant in more schistose rocks. It comprises up to 15% of the whole rock mode and up to 50% of the 'nodules'.

Biotite in these rocks tends to be pale and phlogopitic. Muscovite is ubiquitous and often abundant as a retrograde product of sillimanite (particularly the fibrolite) and of biotite. It is also associated with microcline as a rim against sillimanite or biotite. The K–feldspar seems to have formed as a minor component of some rocks by isochemical metamorphism: other examples show evidence of granitisation. The plagioclase is normally oligoclase. Minor and accessory minerals are the same as those in the non-sillimanitic gneisses and quartzites.

A few rocks contain light pink almandines (Alm_{60-70} Pyr_{17-31}) with concentrations of darker red varieties in quartz–plagioclase segregations. Some sillimanitic bands contain thin graphitic horizons with abundant pyrite. Minor cordierite occurs very rarely.

The sequence of 'nodule' formation can be partially traced, the extent of the process being dependent on the sillimanite concentration and degree of segregation. Incipient nodule development is shown by segregated streaks of sillimanite, elongated in the foliation but grading into the host rocks. Fibrolite is the dominant component of these embryo nodules and sometimes shows helicitic trails, indicating rotational effects during syn-tectonic growth (Starmer 1967).

Further segregation formed discrete quartz–sillimanite 'nodules'. On a macroscopic scale, they have sharp margins, often with a biotite-rich rim, but in thin-section show some gradation with the host rock. The latter frequently contains very little sillimanite. The segregated masses vary from layers, rods and discrete lensoids (1–20 cm in length) to less common small spheroids and irregular masses (a few cm across); they can comprise up to 25% of the total rock mass, the lensoids and spheroids sometimes showing a regular distribution.

Internally, the 'nodules' may contain up to 50% sillimanite as fibrolite and acicular crystals, the latter being particularly concentrated at their margins and as inclusions in quartz. Retrograde muscovite is always present in the nodules and minor or accessory biotite, apatite and magnetite may also occur. An internal fabric may result from orientation of sillimanite, mimetic muscovite (and biotite) and the invariable flattening of quartz.

The rods and layers normally lie in the host-rock foliation, but lensoids are occasionally irregularly orientated or in a parallel arrangement oblique to the matrix fabric. Randomly orientated lensoids are often associated with spheroids and irregular segregations. In some cases, where the 'nodules' and segregations are not concordant to the rock foliation, the latter may traverse them as isolated biotite and muscovite flakes. Commonly, the lensoids occur in the foliation, but may then have internal S-planes showing a slight discordance (usually of 10–15°).

The rods and layers are discontinuous and some show a ductile shear and boudinage: in other cases, contiguous (or almost contiguous) lensoids show formation either by boudinage or by necking and thinning. Similar effects are seen in early, sub-isoclinal minor folds, where lensoids are often elongate in 'a' and may coalesce in the 'ab' plane. Occasionally nodules have broken from segregated layers in sub-isoclinal shear folds, particularly in crenulated, schistose rocks. Many lensoid nodules, however, show no sign of derivation from a more continuous layer or rod. Very late, concentric minor folds have normally deformed the nodule orientations, but in isolated cases (e.g. at Bosvik, Fig. 2), there has been a re-alignment parallel to the axial planes.

The genesis of nodular rocks has been discussed by numerous authors, and Losert (1968) summarised fifteen suggested origins. He considered that they developed late in the metamorphic history of an area by localised 'de-alkalization' and breakdown of biotites and feldspars. It is diffecult to envisage this process reaching the same degree of completion throughout major bands of the Bamble Series.

Recently, the genesis of nodular rocks in the Bamble Series has been considered by Elliott & Morton (1965) and Macaudière & Touret (1969). Elliot & Morton suggested that the nodules formed by boudinage during two fold phases in a series of arkoses and impure sandstones with ribs of high-alumina quartzite. Metamorphism during the second fold phase converted these lithologies to nodular rocks. Within the present area, their F_2 (N–S) folding is a late concentric phase developed only sporadically as minor structures after the main Sveconorwegian tectonism. This process therefore does not seem to explain the field relations or extent of these rocks.

Macaudière & Touret (1969) invoked a process of 'tectonic fibrolitisation' with de-alkalisation in shear zones after the metamorphic climax; a 'dry retromorphosis' paradoxically increasing the grade of metamorphism. The present author has found no evidence of any shear fabric of this type and considers sillimanite to represent the climactic metamorphic conditions with retrograde effects producing muscovite.

It is suggested that the nodular rocks represent part of the original sedimentary sequence. Their distribution and gradation with normal sillimanitic rocks and other metasediments suggest that they were not produced by regional metasomatism. Their early development involved the local ionic mobility of metamorphic differentiation under Upper Amphibolite facies conditions. Completion of this process produced relatively competent quartz–sillimanite lensoids, rods and layers, some of which underwent ductile shear. This development probably started in the Svecofennian orogeny and continued during the Sveconorwegian Regeneration, when the ductile shearing and boudinage were more extensive.

Initial development of 'nodules' by segregation is supported by their non-persistence along the strike, even when the quartz–sillimanite content is sufficiently high. The helicitic, rotational fabrics of some embryo nodules are also evidence of a syn-tectonic segregation origin. Nodules are particularly well-developed around hinge zones of major folds east of Søndeled (Fig. 1) where ductile shearing reached a maximum around the western end of the Levang dome. They are also well-developed in the Søndeled–Akland–Bosvik area (Figs. 2 and 4) where major folds underwent later tightening and bending around the hyperite masses of Barmen and Avreid.

In both nodular and non-nodular rocks, the form of sillimanite as trails of acicular crystals and fibrolite indicates growth during strong, regional tectonism and metamorphism at Upper Amphibolite facies grade. Development during two orogenies would explain the discordance of some nodules (and more commonly their internal fabrics) with the host-rock foliation.

Mixed gneisses

'Mixed gneisses' occur locally throughout the area, but are only developed extensively in the west (Figs. 2 & 4). They consist of rapid alternations of quartzites, variable biotite–quartz–plagioclase gneisses, biotite schists, biotite–hornblende–plagioclase schists and thin amphibolites. They are variably granitised and some bands of granitic gneiss occur in the sequence. It is perhaps noteworthy that sillimanitic layers are generally rare.

Individual lithologies usually vary from a few centimetres to about 2 metres in thickness, forming distinctive banded gneisses in places.

Thin concordant amphibolites

Thin, concordant amphibolites occur sporadically throughout the supracrustals, rarely forming repeated layers except in the western tract of 'mixed gneisses'. They are usually a few centimetres in thickness (only occasionally exceeding a metre) and are too small to be shown on Figs. 2 & 4. The majority are thought to be intercalated basic effusives, but some of the thicker bands rarely show minor intrusive features with bifurcations and apparent xenoliths. In the Granulite facies rocks south of Risør (Starmer 1972b) they seem to have been dominantly intrusive at these deeper crustal levels.

The amphibolites have essential hornblende and plagioclase (An_{30-48}), with minor quartz, biotite and rarer diopside–augite and hypersthene. Some are garnetiferous. Retrogressed specimens have actinolitic hornblende, chlorite and epidote. Accessories include magnetite, ilmenite, sphene, apatite and zircon. Pyroxenes, which are rare, always show strong retrogression to hornblende.

Actinolitic gneisses

These rocks form distinctive layers and bands characterised by the abundance of green actinolitic amphibole. They are usually actinolite/ actinolitic hornblende–phlogopite/phlogopitic biotite–quartz–plagioclase schists and gneisses, with some layers containing little or no quartz, plagioclase or the retrograde mica. Ocassionally interlayered phlogopite (or phlogopitic–biotite) schists occur, containing a little quartz and plagioclase (An_{26-34}).

The main belt of these rocks is surrounded by amphibolitised hyperites between Akland and Molandsvatn (Figs. 2 and 4) with smaller, sporadic developments northwards along the strike (i.e. W and NW of Søndeled). They are also developed in supracrustal enclaves within the hyperites – e.g. at the north end of Lindlandsvann (Fig. 4), where they are associated with orthoamphibole rocks. In the main belt, they seem to have formed by development of actinolitic amphiboles in variable biotite–quartz–plagioclase–hornblende gneisses and thin amphibolites, a sequence tending towards the more diverse 'mixed gneisses' immediately to the west (Fig. 2).

The actinolite or actinolitic hornblende is often nematoblastic, usually up to 0.5 cm in length but reaching 1.5 cm in almost monomineralic segregations. It sometimes poikiloblastically includes quartz-blebs, remnant resorbed hornblende and plagioclase (An_{46-60} in gneisses and An_{33-41} in rocks obviously derived from amphibolite). Discrete plagioclase crystals vary from An_{28-30} in the gneisses to An_{34-55} in altered amphibolites. Complete sericitisation or saussuritisation of all types of plagioclase is common and pale green epidote has often formed.

Phlogopite or phlogopitic biotite is retrograde from amphibole, normally forming strongly lepidoblastic fabrics, but occasionally showing a random orientation. Coarse, segregated layers can be monomineralic but most rocks contain zones in which retrogression has not occurred. This variable alteration reveals that original amphibolites show a metamorphic convergence with the gneisses, when extreme retrogressions produce actinolite–phlogopite (phlogopitic biotite)–plagioclase (An_{28-47})-quartz gneisses.

Discrete hornblende crystals remain in some rocks which show a sequence of changes: hornblende initially became paler, forming actinolitic hornblende followed by the growth of new (often porphyroblastic or poikiloblastic) actinolitic hornblende and actinolite.

Late syn-metamorphic segregations of quartz and plagioclase contain all the above minerals and often show rotated amphibole poikiloblasts with helicitic trails of retrograde micas. This suggests that all the amphibole and mica growths occurred during the main metamorphic episode. Late developments of microcline in some rocks were associated with granitisation. The actinolitic rocks were deformed by very late minor concentric folds.

Accessories in the actinolitic gneisses include ubiquitous magnetite, usually accompanied by sphene and sometimes zircon, apatite and ilmenite. Sporadically, anthophyllite has developed from actinolite, particularly in the coarse segregations.

The origin of these rocks is not fully understood; they seem to have formed from supracrustal gneisses and thin, concordant amphibolites, by porphyroblastic growth of actinolitic amphiboles. In some rocks, the amphibole recrystallisations involved an increase in Si and decrease in Al with an accompanying increase in Mg/Fe ratios. It is possible that some form of metasomatism was associated with an homogenisation of this series of differing bands.

The granitic rocks

These rocks range in composition from granitic (sensu stricto) to quartz monzonitic, granodioritic and adamellitic; they also grade into granitised supracrustals and granitic pegmatite segregations. Within this region (Fig. 1) they are concentrated in certain zones, particularly the Levang dome and smaller bodies to the west, the migmatite zones west of Søndeled and along the Risør peninsula and small domes at Akland, Molandsvatnet and east of Søndeled. Locally the rocks may be massive, but normally they have a biotite, or biotite–hornblende foliation. Remnant supracrustals within them show varying degrees of granitisation and are usually concordant to the granite gneiss foliation. The granitic rock itself is heterogeneous, showing rapid modal variations. Banded migmatites occur in some places.

The lithologies consist of essential quartz–plagioclase–microcline (± biotite) often with minor muscovite or hornblende and accessory magnetite or titanomagnetite. The plagioclase (An_{12-48}) varies from the andesine of assimilated material to newly generated oligoclase. Microcline is frequently replacive and perthitic. Minor or accessory minerals (e.g. sillimanite, garnet) often reflect the pre-granitic parent rock. Other accessories include apatite, zircon and pyrite and retrogressions have formed chlorite and epidote.

The granitic rocks were formed initially by an uprise of metasomatic activity, migma, and small amounts of magma, probably emanating from anatexis at deeper levels. The initial activity was restricted to certain areas by a structural control extending to some depth in the crust. It caused little disruption of surrounding structures and the boundary of granitised zones is often, paradoxically, quite sharp, although minor effects are observed around them. Any isochemical metamorphism of suitable supracrustals to form granitic gneiss was of very limited extent; there is little evidence of this process outside the main granitic zones and it would not account for the restriction of granitic gneiss to the centre of domes.

The granitic rocks are thought to have formed in the Svecofennian orogeny, with some regeneration in the subsequent Sveconorwegian event. This is supported by radiometric data from the Levang dome (discussed later). The Sveconorwegian 'hyperites' intruded pre-existing migmatites, which were subsequently remobilised with break-up of thin amphibolite dykes within them and injection into larger pluton margins (Starmer 1969a). The Sveconorwegian remobilisation left small patches of non-foliate granite, but more significantly caused the extensive development of late granitic pegmatites. Johnson (1976) reports that the western tongue of granite on the Levang dome (Fig. 1) probably represents material remobilised at this time.

The pegmatites

These form bodies ranging from small secretion pegmatites to veins and large discordant masses. A crude zonation is shown by some veins and larger bodies.

Two genetic types can be distinguished (Starmer 1969a): 'granitic pegmatites', ranging from granitic to granodioritic compositions, and 'plagioclase-rich pegmatites', which usually also contain quartz and biotite (± hornblende).

The granitic pegmatites were late hydrous phases associated with the Sveconorwegian regeneration of granitic rocks. They vary from secretion patches, in granitic rocks, to transgressive veins and large discordant bodies cutting all lithologies. They are particularly concentrated around hyperite bodies which formed stable areas of low stress pressure to which fluids migrated.

The plagioclase-rich pegmatites were generated in more mafic rocks by segregation processes during the Sveconorwegian metamorphism. They formed as secretions in the supracrustal gneisses and amphibolites and were mobilised as larger veins and discordant bodies. The mineralogy of smaller bodies is often representative of their parent rocks and they may contain sillimanite, almandine and muscovite.

Granitic pegmatites were essentially post-tectonic, only the small secretions showing any deformation. The larger plagioclase-rich pegmatites were usually not deformed, but some of the earlier secretions were synkinematic and were sometimes intensely folded. Granitic pegmatites may cut all the plagioclase-rich types.

The 'hyperites' (coronites, metagabbros and amphibolites)

A series of gabbroic intrusives consolidated largely prior to the Sveconorwegian metamorphism. The intrusions within this area have been previously considered in detail (Starmer 1969b) and their importance in the present context is as a time horizon. They form plutons, dykes and sheets which cut or partially follow previous structures in supracrustals and migmatites. The subsequent Sveconorwegian orogeny amphibolitised them and caused marginal deformation in and around them. Remobilised granitic material sometimes cut their amphibolitised margins and late granitic pegmatites were concentrated around larger hyperite bodies.

The hyperites were intruded over a period of time and varied from troctolitic types to olivine–norites, olivine gabbros and olivine-free gabbros. Several phases of injection are seen in some bodies, often as coronite dykes a few metres wide, with finer grained margins and perpendicular plagioclase growths against the enclosing, earlier coronite.

The amphibolitisation (which also developed plagioclase-rich pegmatite segregations) left coronite cores in the larger bodies. Glacial erosion often removed their amphibolitised roof down to the level of transition between coronite and amphibolite, where patches of one occur in the other. The extent of the amphibolitisation was also dependent on the availability of H_2O and was therefore somewhat irregular, but these bodies have a general three-dimensional form of coronite core and amphibolite margins.

The coronites have retained their primary igneous minerals surrounded by corona growths. The 'metagabbros' (Figs. 2 and 4) have retained a coarse igneous texture but have an amphibolite mineralogy: the mafics have been replaced by hornblende and labradorite laths pseudomorphed partially or totally by granoblastic andesine. The amphibolites may be massive or foliate with essential hornblende and andesine in subequal proportions, commonly accompanied by biotite, quartz and almandine. Rarely, metamorphic diopside–augite and/or hypersthene (partially altered to hornblende) reflect a local paucity of H_2O during metamorphism. A few amphibolites have developed minor cummingtonite and some late growths of gedrite (± cummingtonite).

The orthoamphibole rocks

Orthoamphibole rocks occur as lenticular masses (normally a few metres, but sometimes tens of metres in length) elongated in the foliation of supracrustals and amphibolites. They are coarse-grained (with crystals occasionally reaching 10 cm length) with a random, or radiating, crystal growth developed after the main tectonism. Rarely they are folded by late concentric folds (Starmer 1967). They grade into the surrounding rocks, but themselves consist of anthophyllite or gedrite with pinitised cordierite, quartz, phlogopite and occasionally remnant plagioclase (An_{15-47}). Accessories include rutile, sphene, pyrite, zircon, muscovite, apatite, haematite and rarely titanomagnetite. Rutile is a common accessory in orthoamphibole-bearing rocks due to the lack of accommodation of appreciable Ti in any of the major mineral phases. Cummingtonite is a major component of some bodies, particularly those developed in amphibolite. The phlogopite (or phlogopitic biotite) is partially formed from original biotite and partially by retrogression of the orthoamphiboles. Muscovite has developed in some rocks from pinitised cordierite.

These lithologies always seem to have formed by replacement of pre-existing metamorphics and, in general, plagioclase has altered to cordierite and clinoamphibole (± biotite) to orthoamphibole. The series of replacements are rather complex and the mineralogy is often complicated since many rocks have been 'frozen' at intermediate stages in their transition.

In amphibolites, hornblende altered to gedrite (± cummingtonite) or more rarely changed totally to anthophyllite or cummingtonite. Any biotite became phlogopitic and when most of the amphibole had altered, plagioclase and phlogopitic biotite changed to cordierite, with the exsolution of some quartz. It seems that some gedrite may have changed to anthophyllite at this time. Similar growth stages are seen in very biotite-rich metasediments and at their junction with amphibolite. Complete alteration formed gedrite (anthophyllite)-cordierite rocks (± phlogopite, quartz).

In plagioclase-rich metasediments the feldspar first altered to cordierite, with some quartz exsolved. In quartz–plagioclase-rich rocks anthophyllite formed late, or was entirely absent, giving cordierite–quartz–phlogopite schists. The latter contain large cordierite porphyroblasts (or poikiloblasts) and differ from supracrustal gneisses, which developed a little cordierite during isochemical metamorphism (see above). Usually, when biotite was present in any quantity (and particularly if hornblende was also present), the biotite became phlogopitic and then orthoamphibole formed (as anthophyllite or gedrite, rarely with a little cummingtonite). The final product was usually cordierite–anthophyllite rocks, but sometimes gedrite formed in gneisses rich in biotite, hornblende or sillimanite. In the actinolitic gneisses, isolated anthophyllite crystals developed sporadically and more intensive replacements formed lenticular bodies of an anthophyllite–cordierite rock (± quartz, phlogopite). The earliest alterations caused anthophyllite to form from actinolitic amphibole and later any plagioclase present altered to cordierite. If there was little plagioclase present (particularly in actinolite segregations) an almost pure anthophyllite rock resulted. The lenticular bodies often have margins of anthophyllite–actinolite rock, sometimes with segregations of cordierite, quartz, anthophyllite and actinolite. To the north of Lindlandsvann (Fig. 4) supracrustal gneisses, forming an enclave in hyperite, are largely actinolitic, particularly near the numerous anthophyllite–cordierite bodies developed. On the Risør peninsula (Starmer 1967), a gedrite-bearing albitite/oligoclasite is developed between granitic gneiss and amphibolite.

The field relations of the orthoamphibole rocks are critical in determining their origins. They are of very limited distribution, occurring as lenses grading into otherwise normal metasediments and amphibolites. The metasediments do not represent magnesium-rich horizons and contain normal biotite rather than phlogopite. In the present area, orthoamphibole rocks are usually associated with the margins of hyperitic amphibolites, whether at the contact or in adjacent rocks on either side. They seem to be particularly concentrated along certain contacts, and south of Vormlitjern (Fig. 4) five small bodies occur within a particular band of biotite–quartz–plagioclase gneiss adjacent to one thin amphibolite.

The orthoamphiboles and cordierite form coarse, random growths and clearly replace other minerals of markedly different chemistry. All features are therefore suggestive of a metasomatic origin, although the source of the necessary Mg-rich solutions is problematical.

Bugge (1943) suggested that Mg-rich solutions emanated from the hyperites during metamorphism. The present author (Starmer 1967, 1969b) found a decrease in Mg during amphibolitisation of troctolitic and olivine-rich noritic hyperites, but this was less noticeable in the other types. To the west of the Levang dome (Fig. 1) Johnson (1976) has found amphibolitised dykes depleted in Mg relative to their parent hyperite plutons.

Morton et al. (1970) did not always observe a spatial relationship between the orthoamphibole rocks and hyperites. They suggested a formation from chlorite–quartz rocks originally produced by low-grade alteration of basic volcanics. In the present area, orthoamphibole rocks show a metamorphic (metasomatic) convergence from different Upper Amphibolite facies lithologies at the end of the main metamorphism. Although they are not always spatially related to larger hyperite plutons, at present levels of exposure, many are associated with thin dykes and apophyses and others may be related to bodies or extensions at depth.

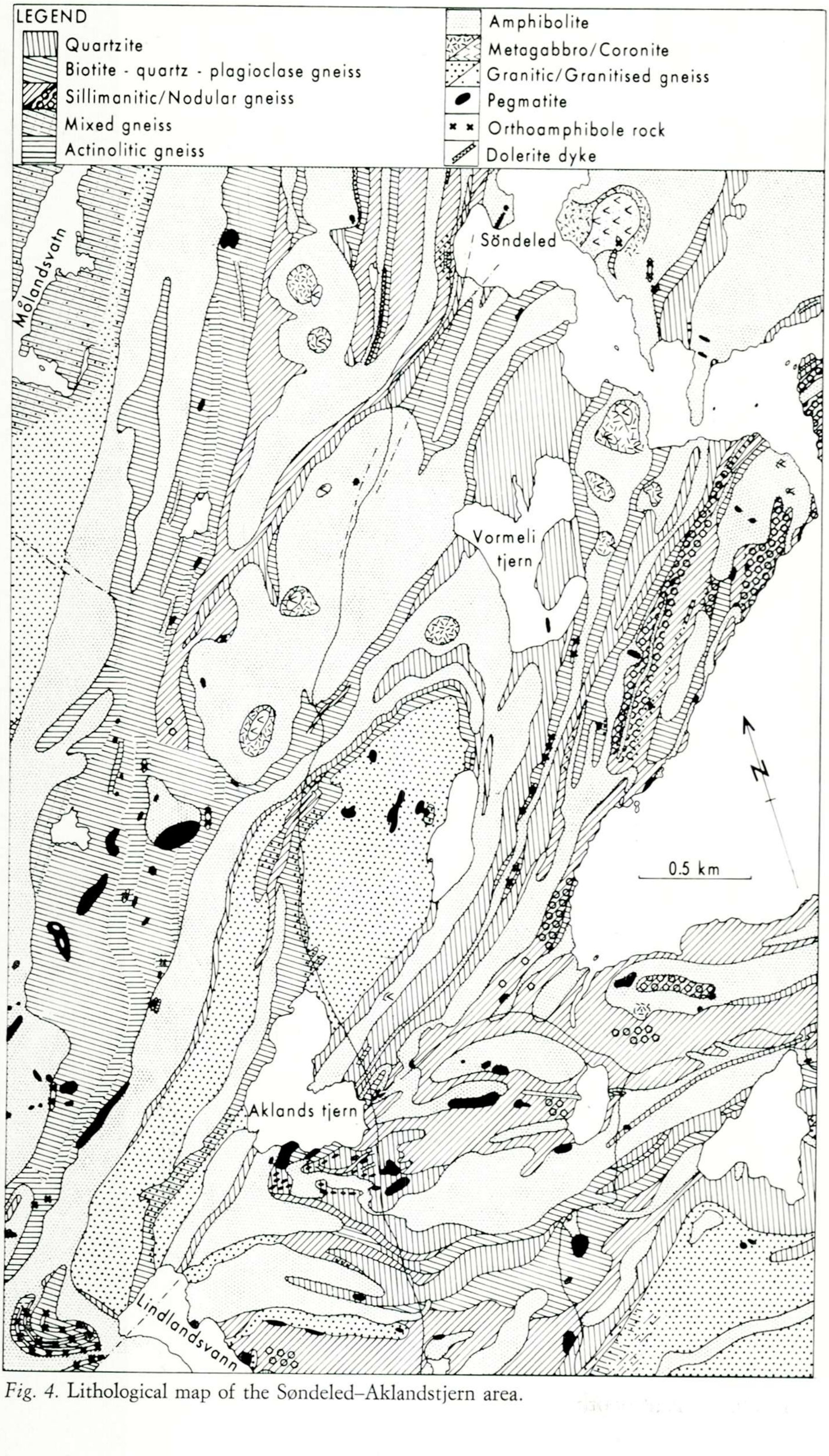

Fig. 4. Lithological map of the Søndeled–Aklandstjern area.

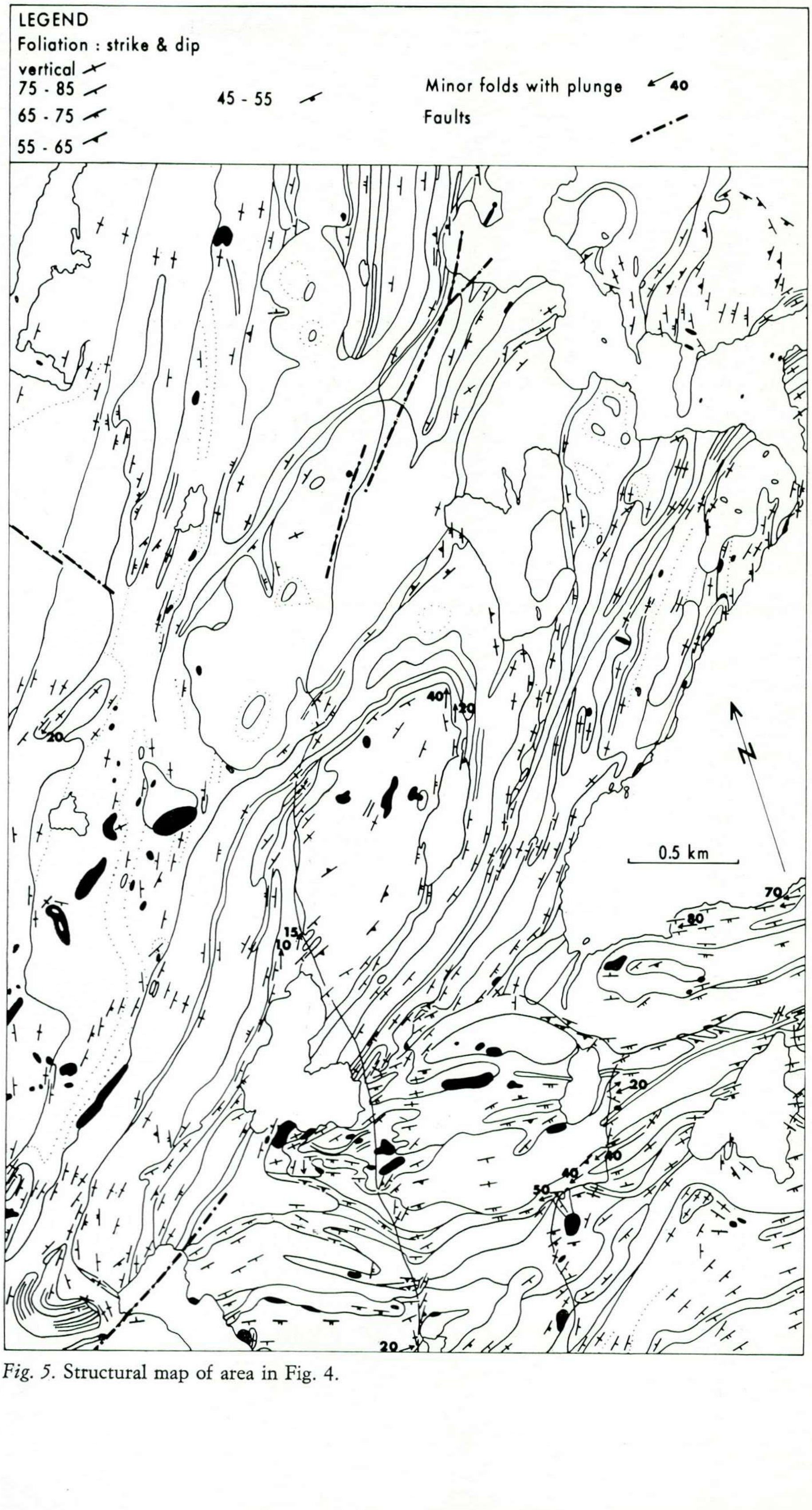

Fig. 5. Structural map of area in Fig. 4.

Dolerite dykes

A number of dolerite dykes occur and three of the largest are shown in the northern part of Fig. 4. They are Phanerozoic in age and (although evidence is not conclusive) are most probably Permian and associated with the Oslofelt activity.

The regional structure

The Levang dome

The Levang dome (Fig. 1) consists of granitic paragneiss containing supracrustal remnants and amphibolites. It was originally described by Hofseth (1942) and has later been surveyed in the east by Elders (1964) and in the west by Johnson (1976). Elders found a number of smaller, second order domes within the eastern part of the structure (Fig. 1). Johnson also reported a number of second order domes and basins and noted local anatexis of the paragneiss, with the southward extension at the dome's western extremity (Fig. 1) probably produced by material remobilised during the Sveconorwegian Regeneration. The Portør peninsula, around the south of the dome (Fig. 1), also contains reworked granitic material (O'Nions & Baadsgaard 1971).

The area southwest of the Levang Dome

Immediately southwest of the Levang Dome (Fig. 1) the rocks have a complex deformation pattern (already briefly discussed) but further southwest, around Søndeled, Akland and Bosvik (Figs. 2 and 3) the structure becomes more ordered. The larger hyperites locally affect surrounding structures, but some have regional effects (e.g. at Avreid and Barmen – Figs. 1–3). Migmatised zones to the west of Søndeled and along the Risør peninsula (Fig. 1) continue southwestwards and coalesce south of Songe.

The major structures in the Søndeled–Akland–Bosvik district (Figs. 2 and 3) were defined at an early stage. They were modified and complicated in two orogenies, with an intervening period of hyperite intrusion (which further increased the existing anisotropy of the region). The major discernible structure is an asymmetric plunging antiform, closing around Akland with limbs striking NNE (to Søndeled) and east (to Bosvik and Risør). The northern part of the Risør peninsula, extending some 8 km east from Bosvik, represents the subsequently modified southeastern limb of this major antiform. Its structure was complicated by smaller scale folding, hyperite intrusions and granitic remobilisations (Starmer 1967, 1969a).

The style and orientation of the major antiform correlate with those of major structures further north in the Bamble Series (Morton et al. 1970; O'Nions & Baadsgaard, Fig. 1, 1970). The Levang dome (Fig. 1) lies between 12 and 30 km NE of the antiform and is suggested to represent another stable remnant on the axial trace of this major structure. The intervening area has a less ordered structural pattern with complex basement-cover tectonics developed east of Søndeled and a major structure on Barmen (along the northern side of the Risør peninsula), produced entirely by deformation around a Sveconorwegian hyperite. This elongate E–W pluton on Barmen developed a steep synform along its length, with complex bending around its western end (Starmer 1967, 1969b). The overall E–W strike of this island and of the Risør peninsula is due to Sveconorwegian compression of the major structure around the Levang dome, with stresses modified by large hyperites at Avreid and Barmen. A marked bend was produced in the western end of the Risør peninsula (Figs. 2 and 3).

On the limbs of the major antiform, smaller scale major folds can be discerned (Figs. 2 and 3). These are second order folds with the first order major antiform forming their median surface (following the terminology of Ramsay 1967) or macroscopic parasitic folds (following the terminology of Turner & Weiss 1963). Particularly obvious on the antiform's northwestern limb are antiformal or domal areas of granitic gneiss at Akland and southwest of Molandsvatn: on the southeastern limb, a large hyperite southwest of Avreid occupies a synform (deformed into a basin). All of these structures were complicated by minor folding, hyperite intrusions, granitic remobilisation and refolding. However, they formed isolated, relatively stable areas, preserving the early structures better than intervening zones which were more severely affected by these processes. They result from a gentle upwarp on a NW–SE axis, running from just southwest of Molandsvatn across the north of Aklandstjern to Avreid. The antiforms at Molandsvatn and Akland were domed and subsequently localised the granitic activity. To the southwest of Avreid (i.e. to the southwest of the upwarp axis) the synform was downwarped into a basin which was later intruded by a large hyperite. (Similar upwarps on NW–SE axes have been noted by Johnson (1976) in the west of the Levang dome.)

Between the isolated stable remnants of early structure (above) only vestiges of complementary folds are now preserved. In the synform between the granitic gneiss domes of Molandsvatn and Akland (and in a tight synform on the southeast side of the latter) resistant hyperite bodies controlled the effects of Sveconorwegian deformations.

The sequence of events

Radiometric data

Recent radiometric work has established the age of certain events. O'Nions & Baadsgaard (1971) produced a Rb–Sr whole rock isochron of 1616 (± 38) m.y. for the northern part of the Levang dome. A later (Sveconorwegian) age of 1167 ± 50 m.y. was obtained from the Portør peninsula, just south of the dome. The rocks of the latter area were granodioritic in contrast to the granitic–adamellitic gneisses of the main dome and were suggested to have been partially melted and reworked. A thermal maximum was defined for the Sveconorwegian metamorphism at 1160–1200 m.y., the major kinematic episode being concluded within 100 m.y., before late pegmatites (the 'granitic' pegmat-

Deformation	Metamorphism	Granitic activity & hyperite intrusions
F_1 minor isoclinal folds. Foliation	Upper Amphibolite	
F_2 (?Svecofennian) major and minor folds (axes generally NE–SW) NW–SE upwarp and doming		
		Granitic activity
F_2 Late pulses		
	Anorogeny	
		Hyperite intrusions
F_3 Main Sveconorwegian deformation	Upper Amphibolite	Granitic remobilisation
	Mid-Amphibolite	Granitic pegmatites
F_4 Minor concentric folds	Lower Amphibolite	
Faulting	Greenschist	

Fig. 6. Sequence of major events.

ites of the present study) consoldiated. The earlier orogenic event was considered to be of Svecofennian age (now termed 'Svecokarelian' by some authors).

Major geological events

A sequence of several major events (Fig. 6) can be discerned from the maps (Figs. 2–5). Granitisation was partly localised in pre-existing antiforms and domes. Later intrusions of basic 'hyperite' were subsequently amphibolitised in an orogeny which deformed the supracrustals around them. Late pegmatite bodies were emplaced after the main tectonism. According to our present state of knowledge, the hyperite intrusions represent the early stages of the Sveconorwegian orogeny.

The earliest recognisable metamorphism of the original supracrustals reached Upper Amphibolite (and local Hornblende Granulite) grade. In many places, early (F_1) minor isoclinal folds are preserved with a penetrative axial planar foliation generally parallel to the lithobanding. Although subsequently deformed, their axial planes seem to have been in the NE–SW quadrants, but not coincident with those of the F_2 deformation which refolded them. Remnant F_1 isoclinal structures are occasionally preserved in granitised rocks.

The F_2 folding was initiated prior to a period of granitic activity, but continued as a series of late pulses after this had finished. The folding had a general NE–SW axial trend and produced the major antiform with its smaller parasitic (major) folds. A gentle upwarp on a NW–SE axis formed the domes at Akland and Molandsvatn. This upwarp may not have been a cross-fold, but rather an activation or emphasis of axial culminations developed during the F_2 compressions and probably intensified by later deformations. In heterogeneous rocks, differential flow due to inhomogeneous compressive strains (differential flattening) along the length of the axial plane, can produce marked fold culminations (Ramsay 1962). Small-scale expressions of this phenomenon caused minor F_2 folds to have variable sub-horizontal to moderate plunges. Interaction with minor F_1 folds has sometimes caused a minor doming of foliation planes (particularly noticeable SW of Aklandstjern and SW of Molandsvatn).

The domes were tectonic in origin and part of the regional supracrustal structure; they were not formed by deformation around central granitic cores. Partially granitised relics sometimes preserve the pre-existing domal structure.

The granitic activity seems to have been a typical Upper Amphibolite facies mixture of metasomatic activity, migma and a little magma, rising in F_2 antiformal and domal areas (e.g. at Akland, Molandsvatn and Levang). Although it continued between some of the later F_2 compressive pulses, it was not the cause of the initial updoming.

These later F_2 pulses foliated the granitic rocks, tightened the folds and elongated the domes. The tightness of F_2 structures was generally inversely proportional to their size; the major antiform was a close to tight structure, but small minor folds became sub-isoclinal. Axial planes had a general NE–SW strike subparallel to the re-orientated lithobanding and S_1 (F_1) foliation, but the axial plunges were variable. Locally, a new foliation was formed, particularly around fold hinges.

A subsequent period of anorogeny ended with the intrusion of gabbroic ('hyperitic') plutons and dykes which cut and partially followed existing structures. Apparent major folds in hyperites are therefore not tectonic hinges. To the south of Risør, hyperites also invaded Granulite facies rocks, which at this time had probably risen to the same level (Starmer 1972b).

Sveconorwegian Upper Amphibolite grade metamorphism amphibolitised the hyperites; isolated developments of pyroxene reflect a paucity of H_2O and local Hornblende Granulite conditions. Plagioclase-rich segregations in amphibolites and supracrustals underwent some deformation but larger, mobilised masses formed essentially post-tectonic pegmatites.

The Sveconorwegian deformations (F_3) acted on a strongly anisotropic complex, with a dominant NE–SW structure stabilised by resistant domes and large hyperites. Compressive stresses were re-orientated and some of the earlier structure was retained and tightened. Compression about the Levang dome caused the Risør peninsula to be bent about its western end to a more east-west alignment. To the east of Søndeled, immediately southwest of the Levang dome (Fig. 1), the same compression squeezed major structures together, with ductile shearing occurring where they were not stabilised by large hyperites. The deformation of some sub-areas was completely controlled by the largest hyperites (e.g. around the Barmen and Avreid bodies, Figs. 1–3). The Avreid mass was intruded into an elongate NE–SW basin, with some discordance on its eastern side. It then controlled the tightening around itself, forming a steep basin structure in its amphibolitised margin and country rocks.

Around most large hyperites, radial compressions produced concentric foliations in their amphibolite margins and country rocks. Elsewhere, the foliation in more elongate hyperites followed that of the enclosing rocks. Locally intense deformations formed minor isoclinal folds in some amphibolites and, on south-east Barmen, developed complex recumbent folds on all scales from a few centimetres to a few metres.

In the domes, the foliation partly follows the major structure and is partly axial planar to the NE–SW elongation. This was due to both late Svecofennian and Sveconorwegian tightening and produced a foliation discordant to the lithobanding around the dome closures.

Local remobilisation and anatexis of granitic material caused some intrusion into hyperite margins (Starmer 1969a, 1972a). The migmatite zones (e.g. the Risør peninsula and the belt north of Molandsvatn) underwent some regeneration, but the discrete domal structures were less affected with only local remobilisations. Sporadic late compressions foliated the early mobilisates, but subsequently non-foliate granitic rocks formed. Late hydrous pegmatitic phases often migrated to stable zones of low stress, particularly around large hyperites and in hinges of recently tightened folds. The assemblages of these granitic pegmatites indicate consolidation under Mid-Amphibolite facies conditions (Starmer 1967).

Very late, open concentric minor folds (F_4) sometimes flattened to a pseudo-similar style, have variable axial directions. Some deform thin pegmatite veins but others are cut by them. The folds resulted from a series of compressive pulses, often only local in effect, but particularly concentrated around large hyperites. Retrogressions and the sporadic development of biotite axial planar fabrics suggest the folding occurred under Lower to Mid-Amphibolite facies conditions. After the main orogeny, brittle faulting under Greenschist conditions was followed by sub–Cambrian peneplanation.

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