Structural development of the Mjelde-Skorelvvatn Zone on Kvaløya, Troms: a metasupracrustal shear belt in the Precambrian West Troms Basement Complex, North Norway

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The Mjelde-Skorelvvatn Zone is one of several metamorphosed and deformed volcanosedimentary belts in the Archaean-Palaeoproterozoic West Troms Basement Complex. The Mjelde-Skorelvvatn Zone is a belt of steeply dipping metasupracrustal rocks (Skorelvvatn Formation) of presumed Palaeoproterozoic age, lying between Archaean (?) anorthositic gneisses (Gråtind Migmatite) and a mafic plutonic complex (Bakkejord Diorite). These units occur on the west limb of a large north-south trending, gently plunging antiform, and all three units share common groups of structures. The structural development of the zone is characterised by three-phase, amphibolite-facies ductile deformation (D_1-D_3) and a phase of brittle/semi-ductile faulting of probable post-Caledonian age. D_1 generated a penetrative foliation (S_1) which is axial-planar to isoclinal folds of variable orientation (F_1). S_1 strikes NNW, dips steeply west and exhibits a moderately plunging stretching lineation (L_1). D_2 involved macroscale folding of S_1 about a gently SSW-plunging axis (F_2). The axis is perpendicular to L_1, suggesting that D_2 may have been a late stage of D_1. D_3 structures are represented by subvertically plunging macrofolds (F_3) with sinistral geometry and a network of moderately to steeply dipping, lateral shear zones (S_3) of dominantly sinistral sense and subordinate dextral sense. The geometric and kinematic relationship between S_3 shear zones suggests that D_3 structures were generated by sinistral transpression with associated minor lateral extrusion. Granitoid pegmatite dykes with a U-Pb titanite age of 1768 ± 4 Ma are shown to have intruded at a late stage of this ductile event, thus constraining a Palaeoproterozoic deformation age. The overall structural character of the Mjelde-Skorelvvatn Zone suggests that it developed during: (i) WSW-ENE crustal shortening (D_{1-2}) and later sinistral transpression (D_3); or (ii) a progressive transpressional event with partitioned east-west crustal contraction (D_1-D_2) and NW-SE lateral shearing (D_3). The proposed kinematic model is comparable with models for other shear zones of the West Troms Basement Complex and is linked with tectonic processes in the Palaeoproterozoic terrains of the northernmost Fennoscandian Shield.

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Introduction

Archaean to Palaeoproterozoic gneisses and intrusive bodies are exposed on coastal islands from southern Lofoten to Vanna, North Norway (Fig. 1). This Precambrian complex is comprised of crustal blocks of migmatites, granitoid rocks and supracrustal belts that were metamorphosed and deformed to variable grades and degrees, and then intruded by granitoid plutons between 1900 and 1700 Ma. Major, steep, NW-SE-striking ductile shear zones segregate these crustal blocks. The Caledonian orogen obscures the relationship between the West Troms Basement Complex and the Fennoscandian Shield to the east. However, and in marked contrast to other Norwegian basement provinces, such as the Western Gneiss Region, the coastal Precambrian rocks in Troms, were subjected to much lesser deformation (reactivation) and resetting of the argon system in late-to post-Caledonian time (e.g. Dallmeyer 1992; Chauvet & Dallmeyer 1992; Olesen et al. 1997).

The main geological components of the West Troms Basement Complex have been defined in many previous studies and are summarised in Zwaan et al. (1998) and Corfu et al. (2003). This paper is, however, the first publication to address the structural complexity of metasupracrustal shear belts in the region, previously examined only in MSc theses (e.g. Pedersen 1997; Armitage 1999). Unresolved questions regarding the metasupracrustal packages are: (i) their relationship to surrounding Archaean migmatites and granitoid complexes, (ii) their modification by Palaeoproterozoic deformation coupled with metamorphism and extensive magmatism, (iii) their possible correlation with Precambrian supracrustal belts in the Fennoscandian Shield further east, and (iv) the nature and extent of Caledonian reworking and post-Caledonian brittle overprinting of the province. This paper contributes to the resolution of these questions by summarising a detailed structural study of the Mjelde-Skorelvvatn Zone (Armitage 1999), which is one of several related

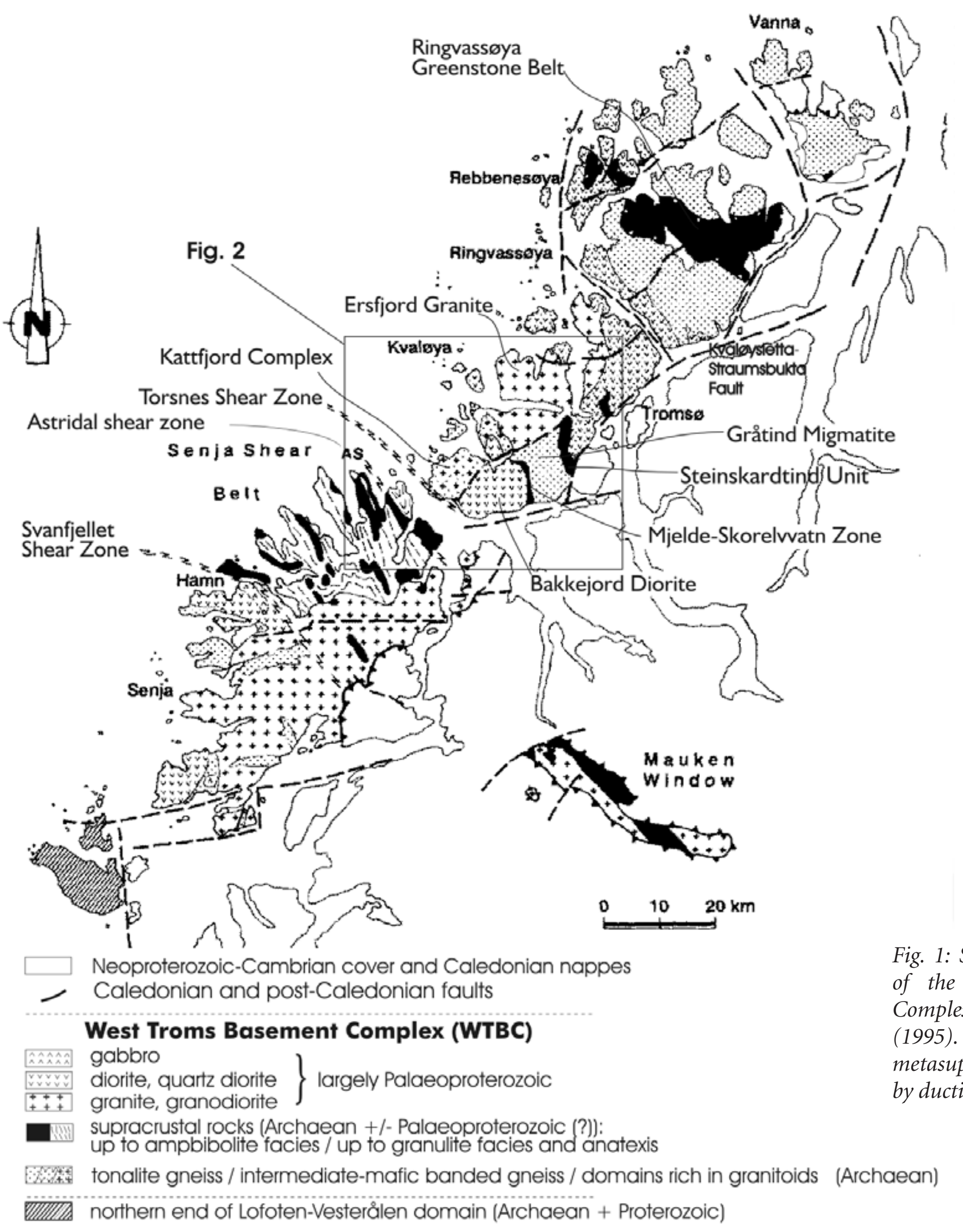

Fig. 1: Simplified geological map of the West Troms Basement Complex, modified after Zwaan (1995). Solid black areas are metasupracrustal inliers bounded by ductile shear zones.

Precambrian shear zones in the region, and by placing a minimum age of 1768 ± 4 Ma on the main shearing event.

Geological setting and previous studies

The West Troms Basement Complex (Fig. 1) is a province of Archaean to Palaeo-proterozoic rocks exposed on Senja, Kvaløya and Ringvassøya (e.g. Zwaan 1992; Corfu et al. 2003). The gneisses of northern Senja and southern Kvaløya occur as megalenses and blocks divided by narrow (<3 km), approximately NW-SE striking ductile shear zones that comprise part of the 30 km-wide Senja Shear Belt (Fig. 1) (Tveten & Zwaan 1993; Zwaan 1995). The Senja Shear Belt is confined in the southwest by the Svanfjellet Shear Zone on Senja (Cumbest 1986; Tveten & Zwaan 1993) and in the northeast by the Torsnes Shear Zone on Kvaløya (Nyheim et al. 1994; Zwaan 1992b). The major shear zones of the Senja Shear Belt are confined to NW-trending, narrow, linear belts of highly strained metasupracrustal rocks commonly associated with ultramafites and intermediate to mafic metavolcanites and mainly quartzitic terrigenous sedimentary rocks. The metasupracrustal rocks have an amphibolite-facies metamorphic signature, locally reaching granulite facies with partial anatexis (Zwaan 1995). Regionally, the Senja Shear Belt is part of the Bothnian-Senja Fault Zone, which is linked to the Svecofennian deformation

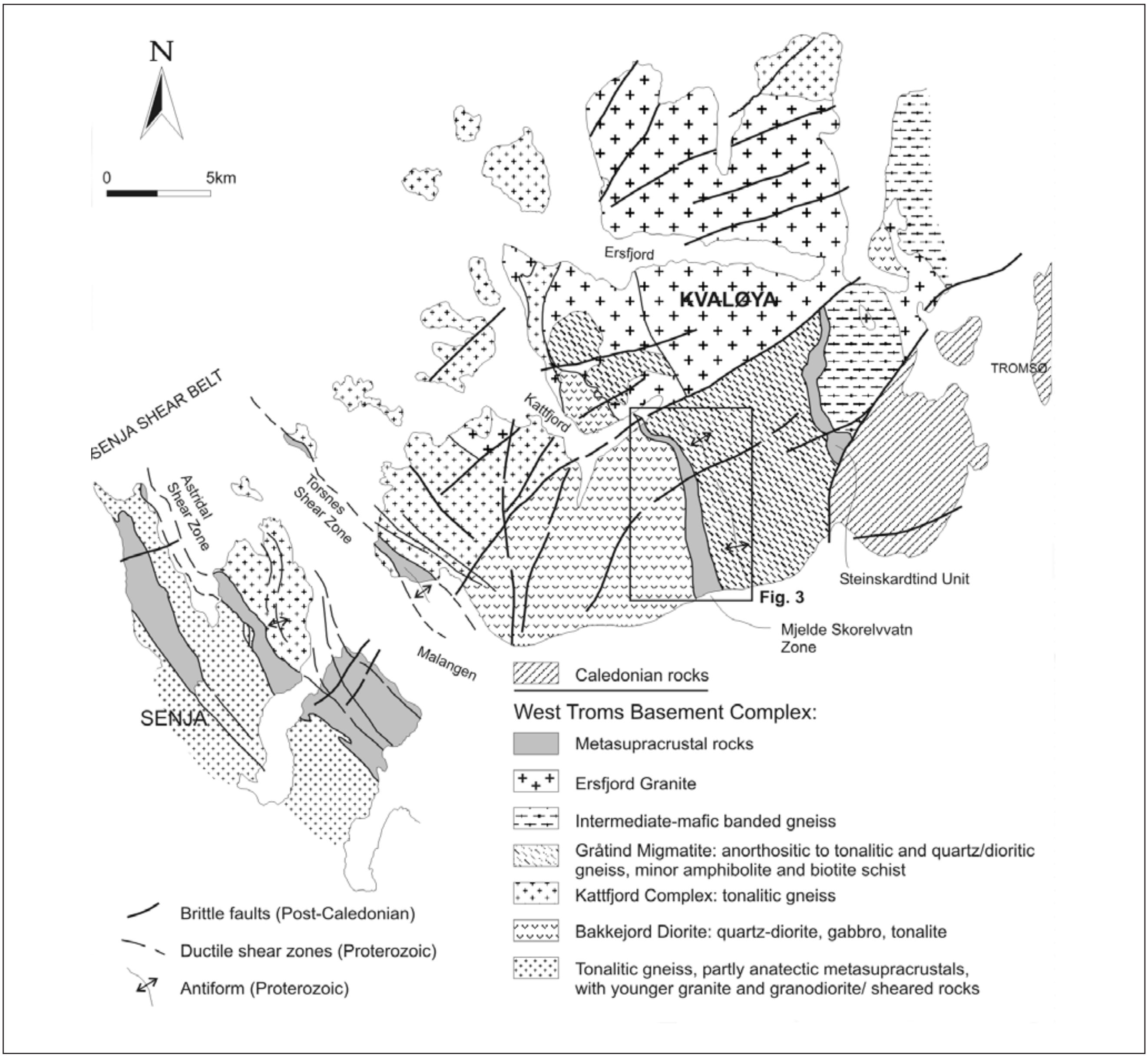

Fig. 2: Geological map of Kvaløya and adjacent areas of the West Troms Basement Complex, modified after Zwaan et al. (1998). Framed area represents the Mjelde-Skorelvvatn Zone.

zones of the Fennoscandian shield by geophysical methods (Henkel 1987). To date, the Svanfjellet Shear Zone is the only recorded Proterozoic shear belt in the West Troms Basement Complex that accommodated weak Caledonian reworking (Dallmeyer 1992).

The geology of Kvaløya (Fig. 2) is described by Landmark (1973), Andresen (1979), Binns (1983b, 1985) and Zwaan (1992a, 1992b). In this paper we apply Zwaan's (op. cit.) terminology for the relevant lithological units of Kvaløya: the strongly foliated Skorelvvatn Formation and Gråtind Migmatite, the weakly deformed Bakkejord Diorite, and the massive Ersfjord Granite (Fig. 2). The dominant strike of the main foliation on Kvaløya is north-south with an eastward dip of 40–50°, but western areas are generally characterised by a subvertical or steep westward dip, as seen in the Mjelde-Skorelvvatn Zone. A similar pattern of opposing dips characterises a NE-SW cross-section of the entire West Troms Basement Complex, expressing a series of large, open, subhorizontal folds. The main foliation is locally rotated into a NW to E-W strike, reflecting a later population of steeply to subvertically plunging folds. Later emplacement of the Ersfjord Granite caused migmatisation of the deformed country rocks up to 3-4 km from its margin, and late-magmatic granitoid pegmatite dykes intruded up to 5-6 km away from the main granite body (Binns 1983b, 1985). The granitoid dykes also transgress common mafic dykes in the Bakkejord Diorite. These mafic dykes, however, do not cut the adjacent Skorelvvatn Formation.

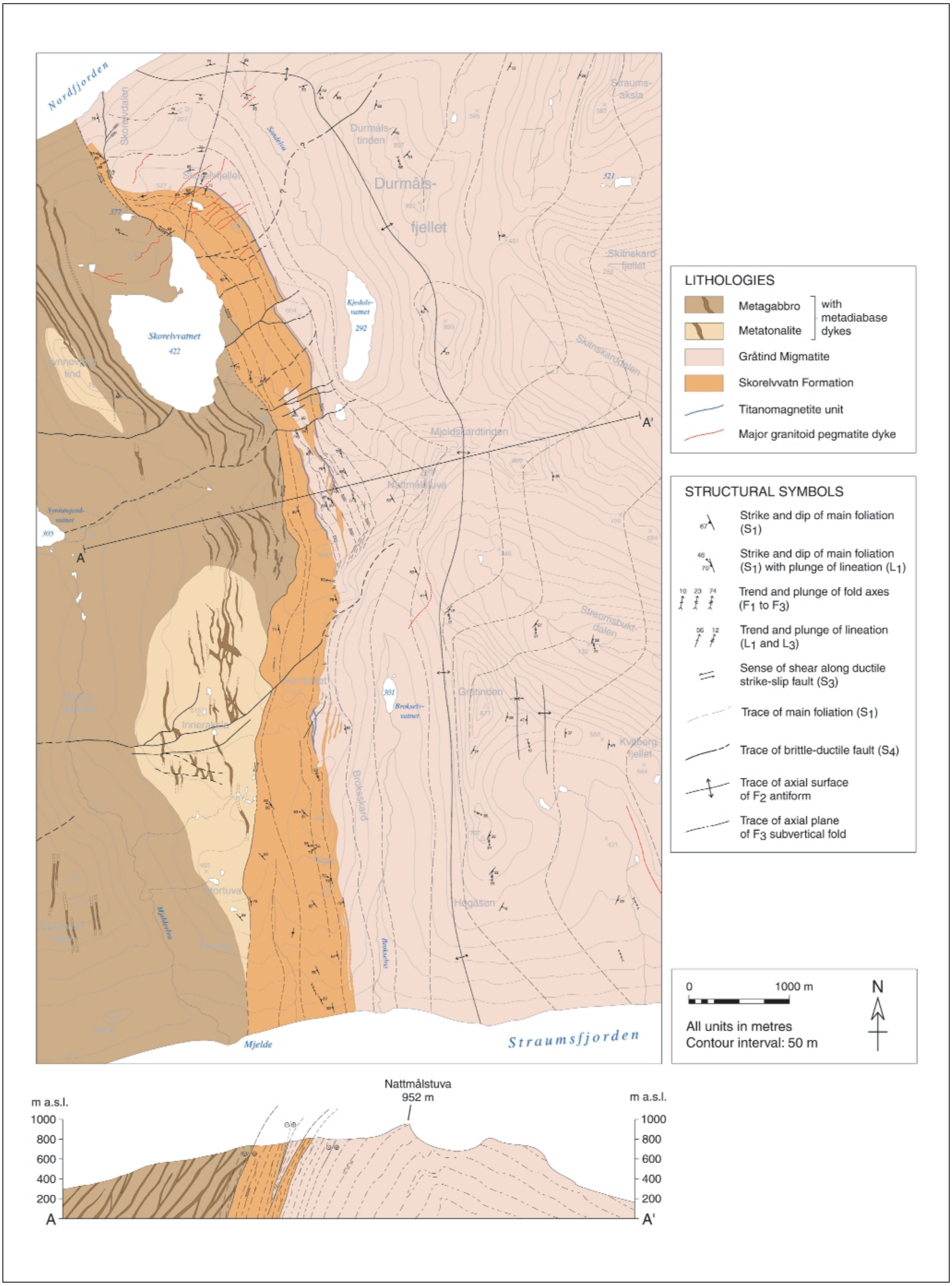

Fig. 3: Geological map and cross-section of the Mjelde-Skorelvvatn Zone, simplified from Armitage (1999).

Important lithological and structural features of the Mjelde-Skorelvvatn Zone were first identified in reconnaissant surveys (Binns 1983b; Zwaan 1992c). The Mjelde-Skorelvvatn Zone is a boundary between the Bakkejord Diorite to the west, which comprises only a weak foliation and abundant metadiabase dykes, and the Gråtind Migmatite to the east, which in contrast is characterised by a strong regional foliation and sparse mafic bands and lenses assumed to be deformed dykes. Although the Mjelde-Skorelvvatn Zone lies outside the Senja Shear Belt, it shares lithological, structural, gravimetric and aeromagnetic similarities with other metasupracrustal belts in the Senja Shear Belt and in other parts of the West Troms Basement Complex.

Rb-Sr whole-rock ages determined for the Ersfjord Granite are 1706 ± 30 Ma (Andresen 1979) and 1779 ± 17 Ma (Romer et al. 1992), and investigation of zircons using the U-Pb system yielded an age of 1796 ± 4 Ma (Corfu et al. 2003). A tonalitic constituent of the Bakkejord Diorite has a poorly constrained Rb-Sr age of 2830 ± 490 Ma (Zwaan & Walker 1996). No age determination has yet been attempted for the Skorelvvatn Formation or Gråtind Migmatite. Radiometric analyses and discussion of age relationships for the main lithological units of Kvaløya can be found in Corfu et al. (2003).

Lithologies

In the study area three main lithological units are exposed on the steeply dipping west limb of a large NNW-SSE trending antiform (Fig. 3): a belt of metasupracrustal rocks (Skorelvvatn Formation) is sandwiched between a plutonic complex to the west (Bakkejord Diorite) and a strongly foliated gneissic unit to the east (Gråtind Migmatite). The boundaries between the three lithological units will be referred to as the Bakkejord-Skorelvvatn contact and the Gråtind-Skorelvvatn contact. The Bakkejord-Skorelvvatn contact is intensely sheared, while the Gråtind-Skorelvvatn contact exhibits relatively minor contact-parallel shear and is transgressed by other shear zones.

The Bakkejord Diorite is a composite intrusive complex comprising metagabbro, metadiorite and metatonalite, cut by a swarm of parallel to anastomosing metadiabase dykes (Fig. 3). It has a very weakly deformed plutonic texture, except for a coarse gneissic fabric near the Bakkejord-Skorelvvatn contact and sporadic pods of highly stretched plagioclase porphyroclasts. The Gråtind Migmatite is an intensively foliated anorthositic to quartz-feldspathic gneiss with sporadic, continuous bands and minor lenses of amphibolite, green serpentinites (retrograde ultramafites) and very coarse-grained biotite schist.

The Skorelvvatn Formation (Binns 1983b; Zwaan 1992a) is a metamorphosed volcanosedimentary sequence dominated by mafic effusives. It consists of coarse-grained, foliated amphibolite intercalated with quartz-ribboned garnet amphibolite, marble, a distinctive 'laminated' actinolite-plagioclase schist and banded psammite. A thin (max. 1.5 m) band of titanomagnetite lies in the Gråtind-Skorelvvatn contact and constitutes a reliable marker horizon. A high, linear aeromagnetic anomaly coincides exactly with the Mjelde-Skorelvvatn Zone and probably indicates a continuation of the titanomagnetite band to considerable depth. Lenses of massive amphibolite occur as coarse-grained to pegmatoidal metagabbro or fine- to medium-grained hornblendite. These may represent relics of mafic volcanics and ultramafic hypabyssal rocks, alternatively, tectonically incorporated lenses of material from the Bakkejord Diorite.

In the Skorelvvatn Formation, the best evidence for sedimentary protoliths is the presence of a continuous marble horizon that grades irregularly into green calc-silicate schist. The central western part of the Skorelvvatn Formation contains fine-grained, slate-like garben schists in contact with lenses of a coarser quartz-garnet amphibolite, both of possible pelitic origin. The more voluminous quartz-garnet amphibolite unit along the east shore of Skorelvvatn is highly distinctive. It contains gedrite as an amphibole phase and, more strikingly, large hornblende 'fans' and flattened garnet crystals up to 10 cm wide. Elsewhere in the Skorelvvatn Formation, indications of volcanic protoliths include apparently pyroclastic fragments in low-strain zones within the amphibolites. Further, the distinctive actinolite-plagioclase schist resembles a finely layered tuff. Stratigraphic polarity indicators, however, have not been found in the Skorelvvatn Formation.

Structural overview

On the basis of dominant structural patterns (D_1 and D_3 elements described below), the map area is divided into three domains (Fig. 4). The north domain is characterised by subvertical macrofolding (F_3) of the main S_1 foliation. The central domain displays sinistral and dextral offset along shear zones (S_3) that cut the main foliation. The south domain is dominated by a planar steep S_1 fabric, with the local exception of an F_3 macrofold at Tverrfjellet (labelled in Fig. 3), which is akin to but less developed than F_3 in the north domain. Areas not incorporated in any domain in Fig. 4 are heavily covered.

The identified structural elements are ascribed to three phases of ductile deformation (Fig. 5). D_1 has generated a penetrative gneissosity and schistosity (S_1) characterised

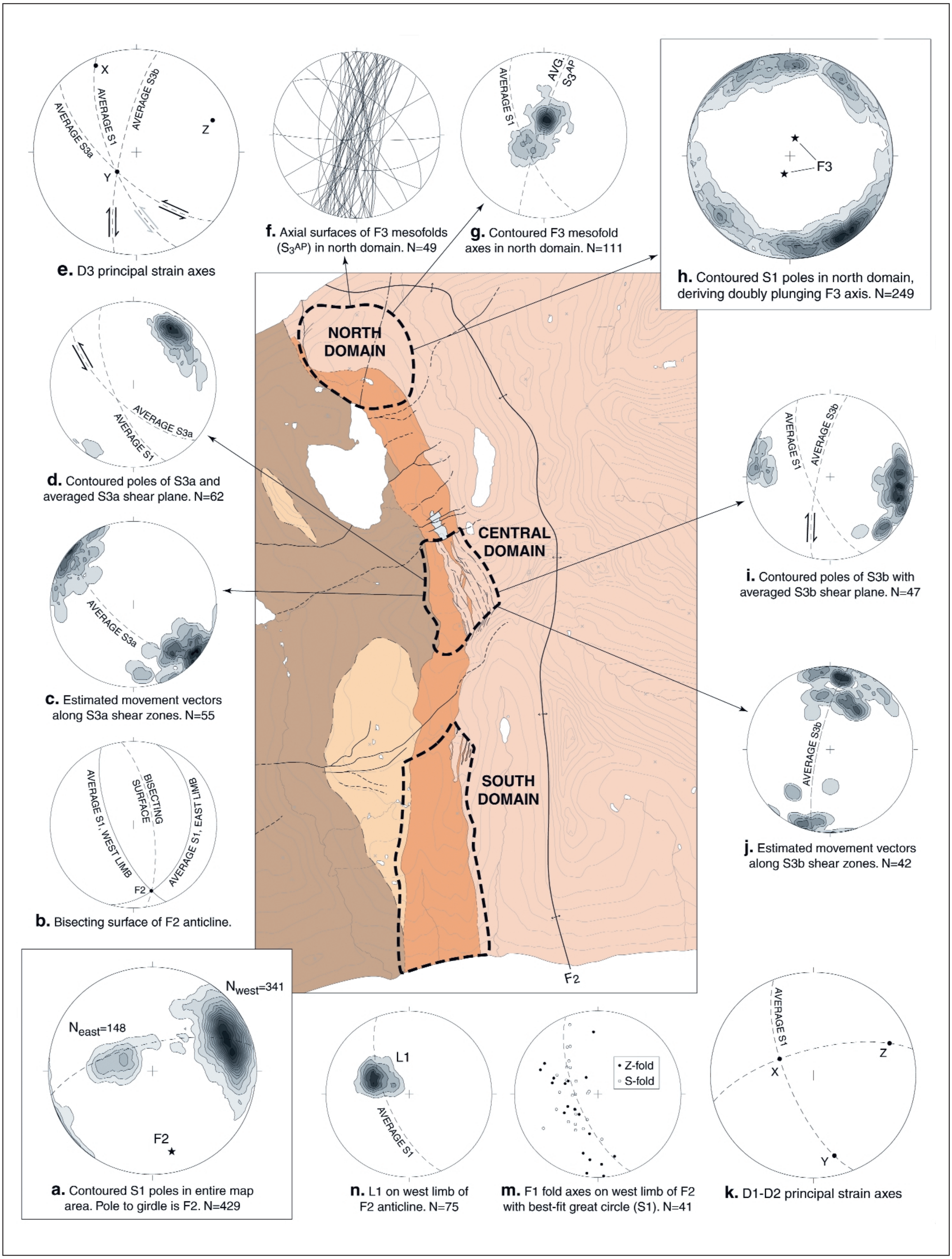

Fig. 4: Division of the Mjelde-Skorelvvatn Zone into structural domains based on the geographical distribution of different D_3 structures. Lower hemisphere, equal-area Schmidt nets with individual captions (a-n) are shown for sets of structures in domains and for specific major structures.

by a mineral assemblage corresponding to middle amphibolite facies. Variably plunging, mesoscale, intrafolial isoclinal folds (F_1) with S_1-parallel axial surfaces are assumed to be coeval with S_1. The main foliation (S_1) is folded by a regional, gently SSE-plunging, open antiform (F_2) with limbs dipping steeply WSW and moderately east. In the west F_2 limb, a WNW-trending, moderately to steeply plunging stretching lineation (L_1) is observed on S_1 surfaces, and is most pronounced in foliated amphibolites of the Skorelvvatn Formation. The Gråtind Migmatite is structurally similar to the Skorelvvatn Formation within at least 400 m of their contact. In the Gråtind Migmatite on the east limb of the F_2 antiform, S_1 strikes N to NE with a gentle to moderate dip. In the Bakkejord Diorite, strongly lineated pods of metadiorite have an identical stretching orientation to the main lineation in the Skorelvvatn Formation and Gråtind Migmatite (L_1).

In the central domain (Fig. 6), S_1 is sinistrally offset by a number of NW-striking, steeply SW-dipping, ductile lateral shear zones (S_{3a}). The titanomagnetite in the Gråtind-Skorelvvatn contact is offset 150–200 m along the two largest S_{3a} shear zones. Smaller macro- and mesoscale shear zones of similar style and orientation are common. S_1 is also dextrally offset by a set of mesoscale, NNE-striking, steeply W-dipping shear zones (S_{3b}). In the north domain, S_1 is folded in a subvertical open macrofold (F_3) towards the Bakkejord-Skorelvvatn contact. The weak foliation in the Bakkejord Diorite near its contact to the Skorelvvatn Formation is locally rotated counterclockwise towards the contact. Granitoid pegmatite dykes are abundant in the northern part of the map area. They cut all lithological contacts and ductile fabrics with the notable exception of D_3 structures. The dykes therefore represent important temporal D_3 strain markers and will be discussed in the continuation.

Fabric elements

First-phase elements: D_1

The first phase of deformation is characterised by: a NNW-striking, moderately to steeply west-dipping, penetrative, metamorphic foliation (S_1); intrafolial, locally transposed, tight to isoclinal folds (F_1); and a stretching lineation (L_1). Within the Skorelvvatn Formation, S_1 is consistently parallel to lithological contacts and to strata-

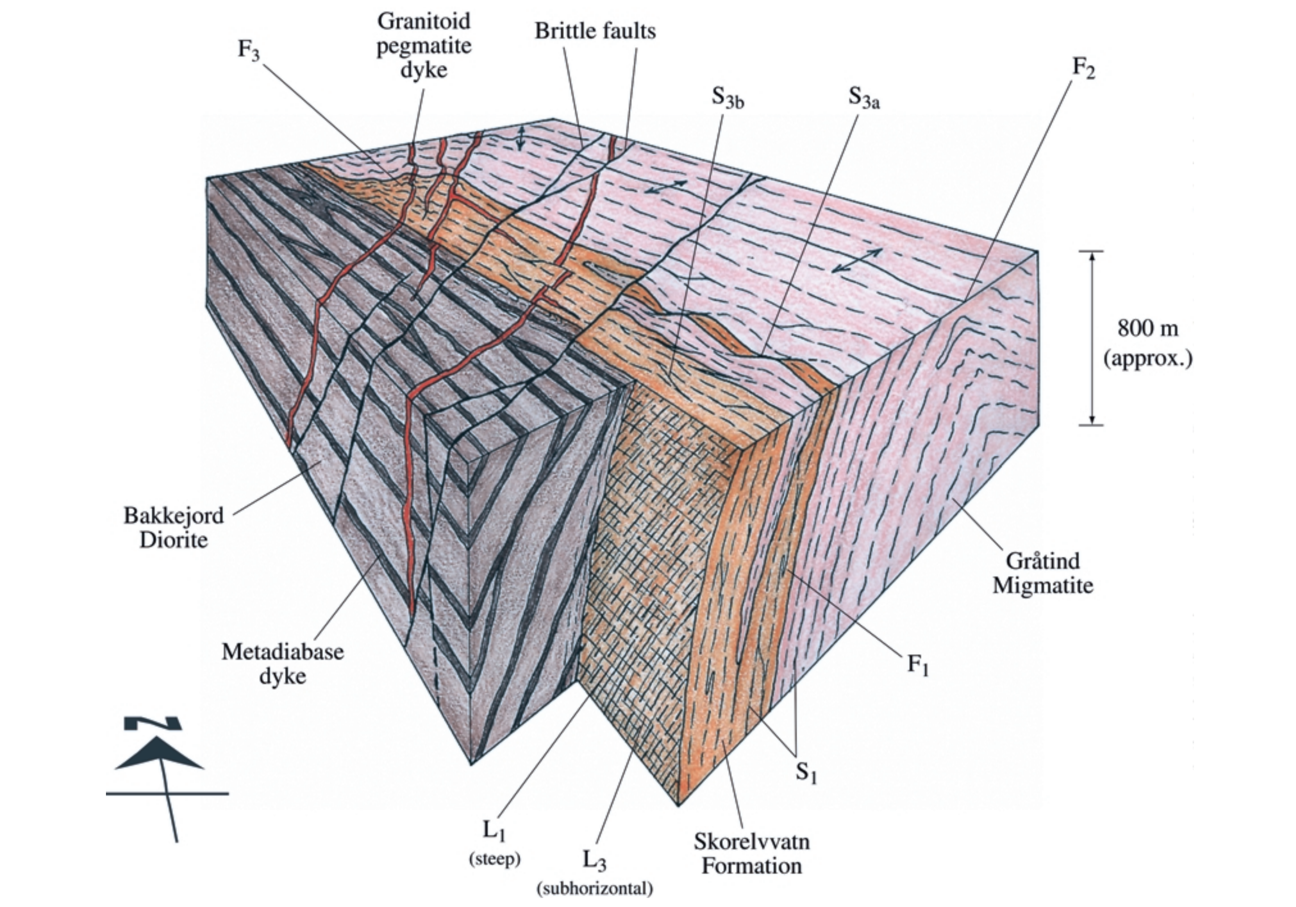

Fig. 5: Three-dimensional block diagram illustrating all the observed macro- and mesoscopic structures of the Mjelde-Skorelvvatn Zone. Note the location of the Skorelvvatn Formation on the west limb of the macroscale F_2 antiform.

bound or lenticular zones of textural variation. Furthermore, S_1 is chiefly defined by hornblende, garnet, biotite and recrystallised quartz and feldspar, signifying lower to middle amphibolite-facies conditions during D_1.

On the west limb of the F_2 antiform, the average strike of S_1 is NNW with a steep WSW dip (Fig. 4a-b). West of Nattmålstuva, the highest peak in the middle of the map area, the strike varies between north and NW, while the strike of S_1 is remarkably constant in the south domain. In the north domain, however, S_1 is rotated towards the Bakkejord-Skorelvvatn contact, expressing a subvertical plunging macrofold (Fig. 4h) in which parasitic mesofolds are common. On the east limb of the F_2 antiform, the strike of S_1 is between NNE and NE with a moderate SE dip.

F_1 folds are represented by mesoscale, intrafolial, tight to isoclinal folds, sometimes transposed (Fig. 7a), and occur as sheath folds in rare instances. The observation that S_1 is consistently parallel to axial surfaces of F_1 folds is reflected by the spread of F_1 axes along the average S_1 plane in a stereonet (Fig. 4m).

In the foliated amphibolite that comprises the bulk of the Skorelvvatn Formation, L_1 is defined by alignment of stretched polycrystalline aggregates of plagioclase and hornblende and locally by stretched volcaniclastic fragments. L_1 has a moderate WNW plunge and lies in the main foliation (Fig. 4n). Highly stretched plagioclase porphyroclasts occur in large isolated pockets in the Bakkejord Diorite (Fig. 7b), producing an L-tectonite with an orientation identical to that of L_1 in the Skorelvvatn Formation. The interplay of the anastomosing S_1 foliation and L_1 stretching lineation in amphibolites of the Skorelvvatn Formation renders an LS-tectonite (generally S>L but L>S in the coarsest varieties of amphibolite).

Second-phase elements: D_2

D_2 elements are represented by open, gently plunging

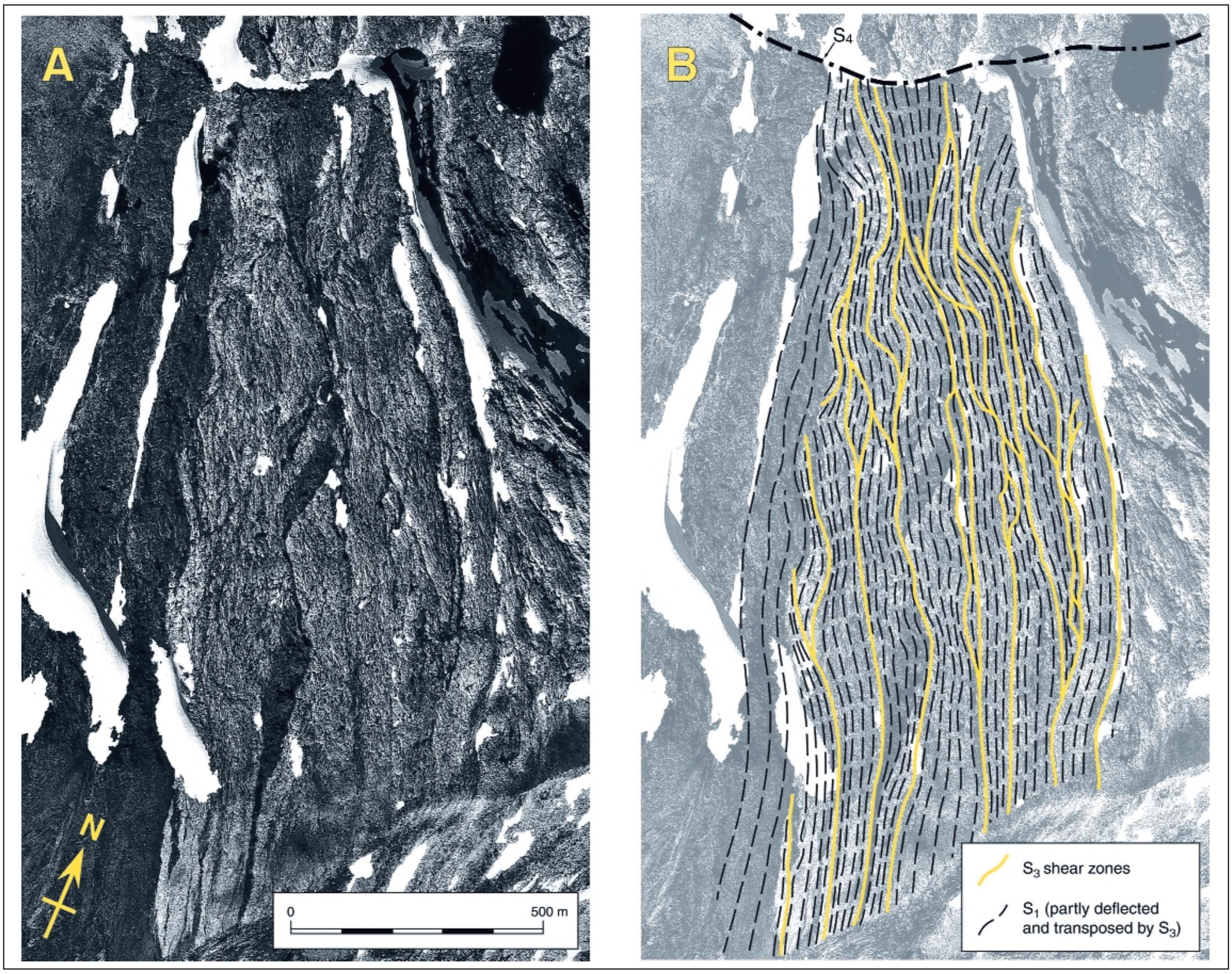

Fig. 6: (A) Aerial photograph of the central domain showing lateral offset along major S3 shear zones. (B) Interpretation of S3 sinistral and reactivated S1 shear zones based on field study. Linear snow-filled hollows in the Skorelvvatn Formation contain calc-silicate schists and marble eroded below the level of the surrounding amphibolites.

folds that fold S_1, in particular the macroscale open antiform that dominates the geological profile and reflects map-scale open folding in the entire West Troms Basement Complex (Figs. 3 and 5). This is a gently SSE-plunging fold with limbs that dip steeply WSW and gently to moderately east. The computed bisecting surface dips ENE (Fig. 4b), showing that the fold is asymmetric and verges WSW. Gently plunging, open mesofolds occur on the east limb of the F_2 antiform. Their weak S-geometry in downplunge view and their similar orientation to the macroscale antiform suggests they are parasitic to the larger structure.

Third-phase elements: D_3

D_3 structures are represented by (see Fig. 5): (i) open, steeply plunging to subvertical F_3 macro- and mesofolds (Fig. 8), (ii) S_1-parallel shear zones that reactivated the main foliation, and (iii) macro- and mesoscale shear zones (S_{3a} and S_{3b}) that cut and deflect the main foliation (Figs. 6 and 8). These structures occur within a zone confined to the Skorelvvatn Formation and the steepest-dipping parts of the Gråtind Migmatite on the west limb of the F_2 macrofold (Figs. 5 and 6). The western limit of D_3 structures appears to be the Bakkejord-Skorelvvatn contact zone. There is no abrupt eastern boundary, but rather a gradual reduction in the frequency of D_3 structures in the Gråtind Migmatite approaching the axial trace of the F_2 antiform.

In the north domain, S_1 is macrofolded around a subvertical axis. The stereographic poles to S_1 here define two girdles and therefore two π-axes, which plunge steeply to subvertically NNE and SW (Fig. 4h). These represent the doubly plunging or undulating axis of a non-cylindrical fold. Similarly at Tverrfjellet in the south domain, S1 in the Gråtind Migmatite is folded by F_3 towards the Gråtind-Skorelvvatn contact.

F_3 mesofolds occur in two main settings: (i) in the F_3 macrofold in the north domain; and (ii) adjacent to S_3 lateral shear zones in the central domain. In a stereonet, the axes of F_3 mesofolds in the north domain define a bimodal clustering (Fig. 4g) that matches the derived π-axes of the non-cylindrical F_3 macrofold (Fig. 4h). In the central domain, mesofolds that occur adjacent to S_3 shear zones are gentle to open, and their surface geometry relative to the shear zones indicates the sense of a lateral component of movement along the shear zones.

S_3 is found in three varieties: (i) an axial-planar cleavage in F_3 meso- and microfolds, (ii) laterally reactivated S_1 planes, and (iii) lateral shear zones that cut S_1. The latter variety of S_3 zones is divided into two sets, named S_{3a} and S_{3b}, each with a distinct orientation and sense of shear (Figs. 4c-e, i-j; 9). For S_{3a} and S_{3b} the alphabetic suffixes denote the major and subordinate set, respectively.

The axial planes of the tightest F_3 mesofolds in amphibolites of the Skorelvvatn Formation are represented by a weak zonal crenulation cleavage defined by aligned hornblende and biotite. The cleavage is steep to vertical and strikes NNE-SSW in most instances (Fig. 4f).

Reactivated S_1 shear zones are represented by continuous zones up to 1.5 m thick and characterised by a phyllonitic fabric (Passchier & Trouw 1998). Where shear planes are observable, some of these zones display a subhorizontal to gently plunging ductile lineation. Viewed down-dip, there are other indicators of lateral shear such as crenulation cleavages and abundant σ- and δ-clasts (Passchier & Simpson 1986). L_1 is obliterated in the most intensively sheared S3 zones. These features suggest that the pre-existing anisotropy (S_1) was reactivated in preferential planes by lateral movement.

Steeply dipping lateral shear zones oblique to S_1 are the characteristic feature of the central domain. S_{3a} strikes NW and exhibits sinistral displacement (Fig. 4d), while S_{3b} strikes NNE and shows dextral displacement (Fig. 4i). Thus, S_{3a} has a counterclockwise acute angle relative to S_1, while S_{3b} has a clockwise acute angle relative to S_1 (Fig 4e). The macroscale S_{3a} shear zones transgress wide areas, including the Gråtind-Skorelvvatn contact, and link up with major reactivated S_1 shear zones to define an anastomosing pattern (Fig. 6).

A few S_3 shear zones display a subhorizontal to gently plunging lineation defined by parallel recrystallisation or stretching of small biotite and hornblende porphyroclasts. All types of S_3 shear zones are characterised by dynamic recrystallisation of hornblende and new growth of biotite, suggesting lower to middle amphibolite-facies conditions. The relative abundance of secondary biotite in S_3 compared with S_1 might be explained by a slightly lower metamorphic grade during D_3 or by more hydrating conditions in D_3 shear zones.

Relationship of dyking to deformation

Metadiabase dykes in the Bakkejord Diorite display the weak S_1 foliation in common with the host rocks, and some metadiabase dykes are clearly truncated along the Bakkejord-Skorelvvatn contact. As these dykes are absent in the Skorelvvatn Formation, they must have been emplaced before D_1 and subsequently truncated by D_1 movement. Their absolute age is not known but would place an upper constraint on the timing of D_1.

Granitoid pegmatite dykes that transgress all lithological contacts in the Mjelde-Skorelvvatn Zone are especially common in the north domain, where they are subparallel to the axial surface of the F_3 macrofold and have intruded along, the axial planes of F_3 parasitic mesofolds (Fig. 8). Some dykes display a ductile

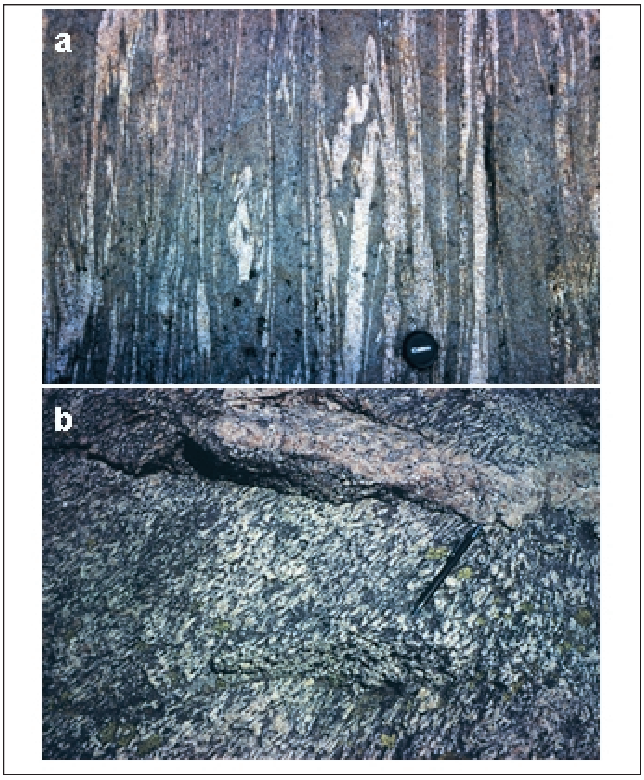

Fig. 7: Photographs of representative D_1 structures: (a) tight to isoclinal F1 folds, partly transposed, in the Gråtind Migmatite near the Gråtind-Skorelvvatn contact, (b) L_1 stretching lineation in the Bakkejord Diorite.

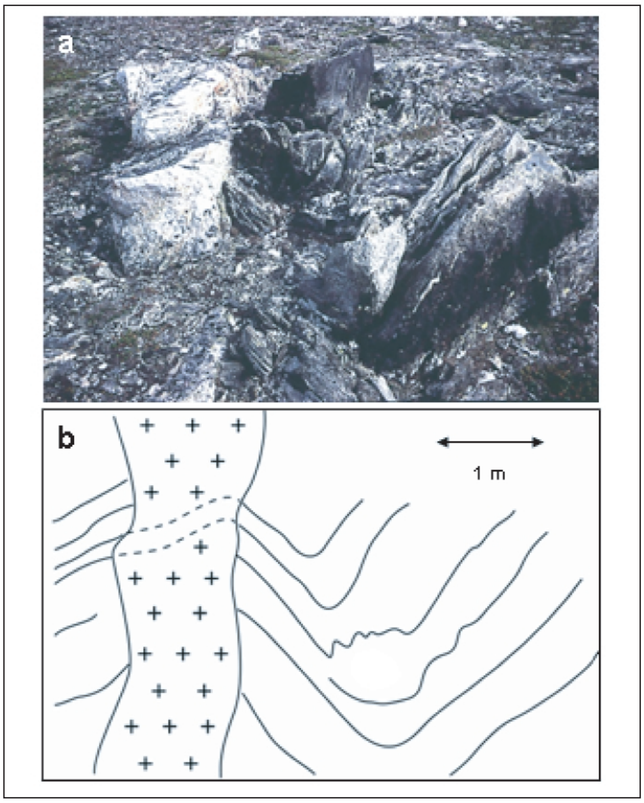

Fig. 8: Photograph (a) and sketched interpretation (b) of a mesoscale F3 fold pair in the northern domain with axial planar crenulation cleavage and a syntectonic granitoid pegmatite dyke slightly sheared in one of the fold hinges.

shear fabric within the hinge zones and limbs of F_3 mesofolds. The shear fabric is parallel to the folded S1 fabric within the folds and is therefore considered to be a result of flexural shear along S_1 during folding. A weak cleavage is locally developed in the pegmatites and is subparallel to the average F_3 axial plane. Further, the dykes show slight sinistral and dextral offset across S_3 ductile shear zones. Some discordant dykes branch parallel to S_1 and are locally folded conformably with the main foliation. These observations indicate that emplacement of the dykes was coeval with the D_3 episode.

Brittle structures

The youngest structures observed are three groups of brittle to brittle-ductile faults, probably, of Caledonian or post-Caledonian in age. Although these structures are beyond the scope of this paper, their effect of brittle faults on the map pattern warrants a brief account. Chloritised, epidotised, slickenlined fault planes were extensively studied during field mapping. The fault zones associated with these slip planes are up to 5 m wide, and the degree of cataclasis increases with width from incipient brecciation to a very fine-grained gouge. Pseudotachylite was not observed but precipitation of haematite is common. Conjugate faults are evident in vertical sections (schematically shown in Fig. 5), especially near the large brittle fault that demarcates the northern limit of the central domain (Figs. 3 and 4). M-plane analysis (Marshak & Mitra 1988) confirms a conjugate set with normal-oblique movement and another, minor set of faults, perpendicular to the conjugate set, that probably formed in the same stress field. The conjugate set has a large normal component of movement and a minor sinistral strike-slip component. Most of the major faults seen in map view are the synthetic element of the conjugate set, striking WSW-ENE with a steep dip to the northwest. Some splaying faults are also evident (Fig. 3).

Kinematic analysis

D_1 strain

As the axes of F_1 folds have highly variable plunges in the S_1 plane, and because the stereographic pattern of F_1 parasitic folds with S- and Z-geometry shows no clustering (Fig. 4m), the folds cannot be used as kinematic indicators. Possible explanations for the variable orientations of F_1 fold axes are: (i) axes were progressively and differentially rotated within S_1 during D_1, or (ii) axes of initially similar orientation were variably rotated by a later phase of deformation (D_3 lateral shear). In the former case, the folds represent non-cylindrical structures that in some instances developed into sheath folds (Bell 1978; Bell & Hammond 1984).

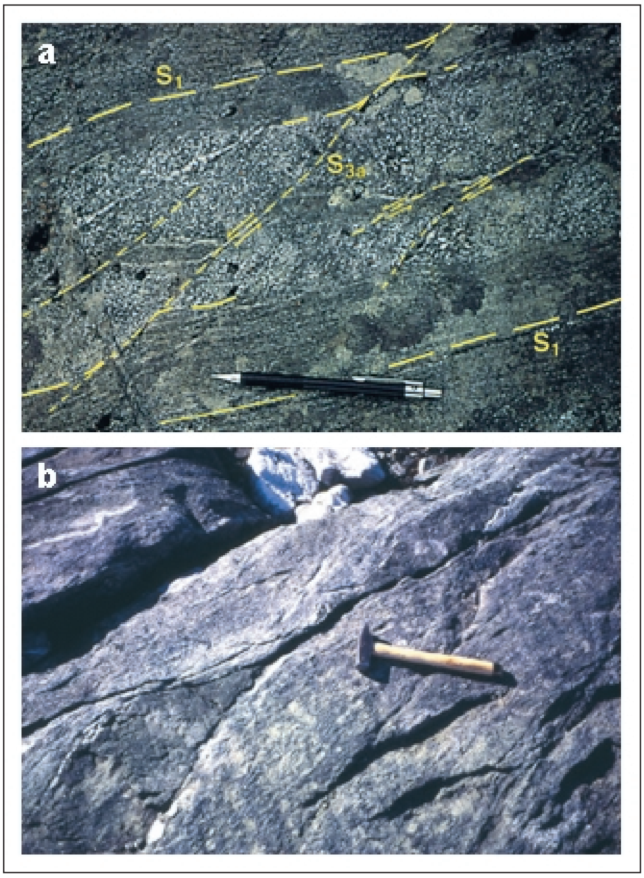

Fig. 9: (a) Sinistral S_{3a} shear zone cutting a dark hornblendite layer in the Skorelvvatn Formation. (b) Small-scale dextral S_{3b} shear zones displacing dark bands in Skorelvvatn amphibolites.

These fold types are present but uncommon in the study area.

Consistent repetition of the titanomagnetite band between the Skorelvvatn Formation and Gråtind Migmatite in the central domain (Fig. 3) suggests that the repetition is tectonic rather than stratigraphic. The repetition can be explained by isoclinal folding and/or thrust wedging during D_1. Such an event could not have occurred later than D_1 because the repeated Gråtind-Skorelvvatn contact is cross-cut by later S_{3a} shear zones. Open macrofolding (D_2) is considered unlikely to have caused the repetition in view of the fairly uniform dip of S_1 across the repeated section. Any difference in geometry between isoclinal folding (consistent S_1) and thrust wedging (some discordance in S_1 across a thrust) would now be difficult to detect as the main foliation has been macrofolded (D_2) and then cut and deflected by lateral shear (D_3).

The SL-tectonites defined by the combined S_1 foliation and L_1 stretching lineations provide more reliable kinematic indicators, particularly the L>S amphibolites. Estimated strain axes for the D_1 event are presented in Fig. 4k. The axis of greatest elongation (X) is approximated by the average orientation of L_1. The axis of greatest shortening (Z) is the pole of the average S_1 plane. The intermediate Y-axis is perpendicular to the X and Z axes. Stereographic reconstruction for the D_1 phase suggests an overall WSW-ENE orientation for L_1 (Fig. 4k). This orientation is proven to be almost perpendicular to the strike of S_1 and is an indication of the axis of D_1 contraction in the present compass reference.

D_2 strain

The axis of the macroscale F_2 antiform is assumed to represent the Y-axis of D_2 strain and is evidenced to be virtually normal to L_1 (Fig. 4b, k, n). This geometric relationship implies that antiformal macrofolding could be genetically linked to D_1, i.e. D_2 open subhorizontal folding could be an incremental stage of the D_1 foliation-forming event. Temporal continuity between D_1 and D_2 is further supported by the consideration that the L_1 stretching lineation is too elongated to be a product of open flexure alone (F_2). The significance of the WSW vergence of the macroscale F_2 antiform is questionable, however. Two possibilities are that the direction of D_1-D_2 tectonic transport was WSW, or the west F_2 limb became steeper due to D_3 contraction. Our kinematic model favours the latter scenario.

D_3 strain

D_3 strain can be inferred from the asymmetry of steep to subvertical F_3 folds in the north domain, where one of two F_3 macrofolds is well exposed. The fold has a doubly plunging axis (Fig. 4h), and the narrow angle between the two axes may relate to an undulating vertical to subvertical axis, i.e. a non-cylindrical fold. Alternatively, the fold formed during D_3 sinistral strike-slip shear and was modified slightly by minor D_3 dextral shear. This does not necessarily mean that the sinistral and dextral shearing were temporally separate. If the doubly plunging F_3 axis is a result of dominant sinistral folding and modification by dextral strike-slip movement, the SW-plunging component could be the result of sinistral shear since it corresponds to the axes of open folds adjacent to S_{3a} shear zones. The NNW-plunging component may then reflect dextral modification, corresponding to very open folds adjacent to S_{3b} shear zones.

The total dataset for F_3 mesofold axes parasitic to the F_3 macrofold defines a bimodal clustering in a stereonet (Fig. 4g). One set lies in the average main foliation (S_1) and in the averaged axial surface of F_3 mesofolds (S_3). Therefore, their development was likely controlled by the pre-existing foliation (S_1). The other set of mesofolds lies in the average axial surface of the F_3 macrofold, which suggests a development less dependent of S_1.

The reliability of S_3 shear zones as shear sense indicators depends heavily on their mechanism(s) of formation. Four different models are considered for S_{3a} and S_{3b}: (1) ductile analogies to Riedel shears generated by bulk simple shear (Sylvester & Smith 1976); (2) normal slip crenulation (Dennis & Secor 1987); (3) shear-band cleavage (Williams & Price 1990); and (4) shear bands generated during lateral extrusion (Jones et al. 1997).

(1) S_{3a} shear zones are geometrically and kinematically similar to synthetic Riedel shears. In brittle regimes, synthetic Riedels (R) and antithetic Riedels (R´) usually form a conjugate pair about the resolved maximum compression direction, but the synthetic set can dominate over the antithetic set both in number and magnitude of displacement. However, no analogy to R´ for a sinistral regime has been observed in the study area. Likewise, S_{3b} cannot represent R in a separate dextral regime because it has no antithetic counterpart. The Riedel shear analogy is therefore rejected.

(2) At all scales, S_{3a} shear zones have the same configuration as normal slip crenulations (NSC – Dennis & Secor 1987) in a zone of sinistral simple shear. Normal slip crenulations offset the foliation in a normal (extensional) sense that is synthetic with that of the main zone and is analogous to S-C structures (Berthé et al. 1979; Lister & Snoke 1984; Simpson & Schmid 1983), shear bands (White et al. 1980), and extensional crenulation cleavage (Platt & Vissers 1980). The crenulation compensates movement normal to the shear zone boundary resulting from slip along a foliation oriented at a counterclockwise acute angle to the shear zone movement direction. Where slip occurs along a foliation that is parallel to the shear plane, minor perturbations in the foliation or lithological inhomogeneities might induce crenulations. Also, the development of normal slip crenulations (or – reverse slip crenulation) may inhibit further slip along the foliation.

Geometrically, S_{3b} is simply a mirror image of S_{3a} and may then represent normal slip crenulations in a dextral strike-slip regime. If S_{3a} and S_{3b} represent normal slip crenulations as generated in the manner described by Dennis & Secor (1987), then the D_3 shear plane must have been oriented approximately NW-SE such that the NNW-SSE striking main foliation had a counterclockwise acute angle to the shear plane. Alternatively, S_1 is parallel to the D_3 shear plane and the shear bands are a result of perturbations and inhomogeneities, examples of which are seen along the Bakkejord-Skorelvvatn contact and within the Skorelvvatn Formation in the form of massive mafic lenses.

(3) The development of kinkbands or shear-band cleavage (Williams & Price 1990) when the schistosity is parallel to the shear-zone boundary indicates that the shear zone is transtensional or transpressional, respectively. Conjugate shear bands tell that schistosity-normal shortening was relatively strong, whereas single shear-band sets indicate that schistosity-parallel shear was relatively strong. Shear-band cleavage is analogous to normal slip crenulations (Dennis & Secor 1987), but expresses transpression (oblique convergence) when it develops in a zone where a pre-existing foliation is parallel to the shear plane. It is conceivable that S_{3a} in the Mjelde-Skorelvvatn Zone represents shear-band cleavage in sinistral transpression, in which case S_{3b} can also be interpreted as shear-band cleavage but resulted from dextral transpression. D_{3a} and D_{3b} could then be separate phases of deformation, though not necessarily separated in time because they show exactly the same style of deformation and metamorphic signature. In the continuation, another model will be offered for the synchronous operation of S_{3a} and S_{3b}.

(4) Lateral extrusion has also been called 'axial elongation' (Harland 1971) or 'lateral escape' (Dias & Ribeiro 1994). It is defined as a stretch in the horizontal direction that causes a deformation zone to lengthen in transtension or shorten in transpression relative to the undeformed rocks that comprise the zone margins (Jones et al. 1997). The lateral extrusion model eliminates some of the restrictions imposed by boundary conditions in other transpression models. The model assumes an unconfined transpression or transtension zone in which material may move vertically and horizontally in and out of the deforming zone parallel to its length. The obtuse bisector of shear zones is an approximation of the axis of greatest shortening. Antithetic strike-slip shearing is a kinematic requirement of laterally unconfined transpression, and synchronous shear-sense indicators may reveal opposite senses of movement. S_{3a} and S_{3b} in the Mjelde-Skorelvvatn Zone have opposite senses of movement, and their identical style and metamorphic signature support a synchronous genesis. Synthetic sinistral shear in the Mjelde-Skorelvvatn Zone is substantiated by the predominance in number and magnitude of S_{3a} shear zones and F_3 sinistral folds. The fewer and smaller S_{3b} shear zones are therefore interpreted as the antithetic set.

It is not known whether the main foliation (S_1) is parallel or oblique to the overall D_3 shear plane of the Mjelde-Skorelvvatn Zone. If they are parallel, the presence of secondary discontinuities in the form of S_{3a} suggests transpression, and the presence of S_{3b} as a possible antithetic set implies lateral extrusion (Jones et al. 1997). The obtuse bisector of the shear zone sets is an approximation of the axis of greatest shortening (Z-axis) (Fig. 4e). This axis is oriented approximately E-W in the present compass reference, and E-W shortening across a NNW-SSE oriented boundary involved a component of sinistral shear along the boundary represented by the Skorelvvatn Formation. Observed L_3 lineations in the Mjelde-Skorelvvatn Zone confirm subhorizontal to gently oblique movement along S_{3a} and S_{3b} shear zones (Fig. 4c and 4j).

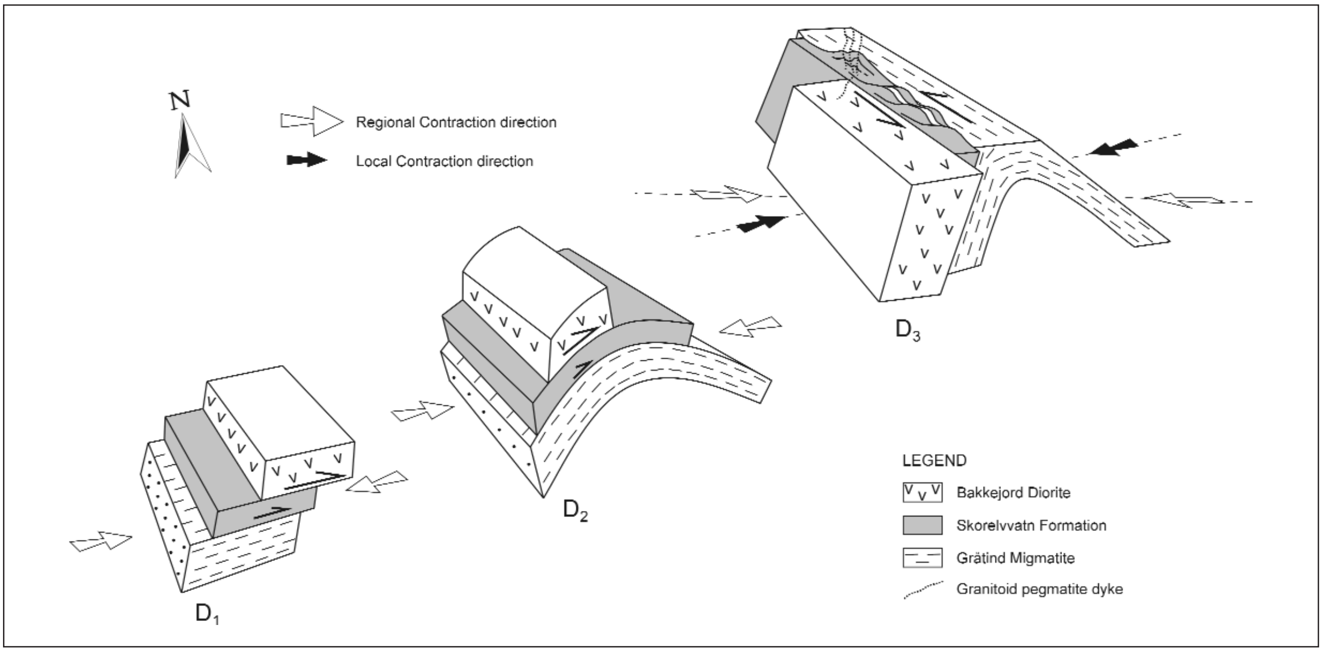

Fig. 10: Simplified three-stage (D_1-D_3) kinematic model for the Mjelde-Skorelvvatn Zone. [D_1] WSW-ENE crustal contraction caused thrusting and generation of F_1 isoclinal folds, main S_1 foliation and L_1 stretching lineation. [D_2] Continued WSW-ENE contraction produced F_2 macroscale antiform. [D_3] Further oblique NW-SE contraction steepened the west limb of the F_2 antiform, which consequently became a favourable locus for lateral shear. The resultant sinistral transpression locally reactivated S_1 and generated major synthetic S_{3a} shear zones, minor antithetic S_{3b} shear zones (as a component of lateral extrusion), steep to subvertical F_3 folds and shallow-plunging to subhorizontal L_3 stretching lineation.

Kinematic models

Two different kinematic models for the Mjelde-Skorelvvatn Zone can be deduced from the data and analysis presented above. One model (Fig. 10) invokes an initial phase (D_1) of approximately WSW-ENE contraction that generated a penetrative foliation (S_1) and associated structures in amphibolite-facies metamorphic conditions. Later WSW-ENE contraction (D_2) resulted in open, NNW-SSE horizontal folding of D_1 structures (F_2). A third stage of deformation (D_3) was approximately NW-SE contraction under the same metamorphic conditions as D_1. The response to D_3 on the steep west limb of the F_2 antiform was sinistral transpression with lateral extrusion, generating a dominant set of synthetic shear bands (S_{3a}) with sinistral macro- and mesofolds, and a subordinate set of antithetic shear bands (S_{3b}) with associated dextral mesofolds.

A second and preferred model is one in which D_1, D_2 and D_3 are increments of a progressive partitioned event in amphibolite-facies conditions, involving WSW-ESE oblique contraction (transpression) across a NNW-SSE-striking boundary represented by the Skorelvvatn Formation. S_1 must have formed at an early stage, as it predates the granitoid dykes that are evidenced to be synchronous with a late stage of D_3. Likewise, the F_2 antiform must have developed before D_3 because it appears in map scale to be refolded by F_3-folds. S_{3a} and S_{3b} shear bands are the synthetic and antithetic expressions, respectively, of lateral extrusion during sinistral transpression. This model is favoured because it is the best explanation for the geometric relationships, temporal relationships and similar metamorphic signatures seen in all ductile (D_1–D_3) elements. The localisation of lateral (strike-slip) shear structures in the Skorelvvatn Formation was likely controlled by its relatively incompetent rheology and by the increasingly steep orientation of its main fabric during progressive transpression.

Age of deformation

A granitoid pegmatite dyke in the northern domain of the map area was sampled for age determination, as it is one of many dykes that show a syntectonic relationship to D_3 structures (Fig. 8). As mentioned in the structural description, where dykes are observed to have intruded along F_3 axial planes, the intrusive lithology shows a ductile fabric resulting from flexural shear in the F_3 hinge zones. The dykes are also deflected and cut by S_3 shear zones. If the dykes were pre-D_3 intrusions, they would not show an association with F_3 axial planes and would not be deformed in F_3 hinge zones alone, but would rather be deformed more extensively by the folding. If they were post-D_3 intrusions, they would not be deformed by S_3 shear zones. The observed relationships and the slight degree of deformation suggest the dykes were emplaced at a late stage of D_3. Further, they provide constraints on the relative timing of sinistral (S_{3a}) and dextral (S_{3b}) shear. If dextral shear had occurred earlier than sinistral shear, the dykes would have been first offset dextrally by S_{3b} and show no relationship to the axial planes of later F_3 sinistral folds in the north domain. Therefore, sinistral shear must have been earlier than or coeval with dextral shear.

Zircons extracted from the selected dyke were unsuitable for age determination, but the earlier of two titanite generations yielded an age of 1768 ± 4 Ma (Corfu et al. 2003), interpreted as a minimum age of crystallisation and therefore a minimum age for D_3. This age is compatible with whole-rock Rb-Sr ages of 1706 ± 30 Ma and 1779 ± 17 Ma previously determined for the parent Ersfjord Granite body (Andresen 1979; Romer et al. 1992) when sources of error are taken into consideration. It is also compatible with a more reliable U-Pb zircon age of 1796 ± 4 Ma determined for the Ersfjord Granite (Corfu et al. 2003). A syntectonic pegmatite dyke at Otervika west of the Mjelde-Skorelvvatn Zone yielded a zircon age of 1774 ± 5 Ma (Corfu et al. op. cit.), which together with the 1768 ± 4 Ma dyke in the Mjelde-Skorelvvatn Zone indicates that lateral shear was then at a waning stage. It is noteworthy that the pegmatite dyke sampled in the Mjelde-Skorelvvatn Zone contains a second generation of titanite dated to 1751 ± 8 Ma (Corfu et al. 2003), assumed to have grown during a late metamorphic and possibly deformational overprint. This late phase corresponds to a titanite population dated to about 1755 Ma in two dykes at Otervika (Corfu et al. op. cit.).

Discussion and regional implications

Binns (1983b) considers the Skorelvvatn Formation to be an *in situ* xenolithic unit in the felsic rocks represented by the Gråtind Migmatite and partly by the Bakkejord Diorite. Whilst this is possible, we consider it unlikely given that the Skorelvvatn Formation lies between two lithological units of evidently different character. Zwaan (1992a) offers three possibilities for the present configuration of the Skorelvvatn Formation: (1) a highly deformed shear zone in the Bakkejord Diorite, which we reject in view of the interpreted supracrustal origin of the Skorelvvatn Formation. (2) An infolded supracrustal sequence with a basement comprised of both the Bakkejord Diorite and Gråtind Migmatite. This is similar to Binns' (1983b) argument above. Again we consider it unlikely because the Skorelvvatn Formation is flanked by two distinct lithological units, and the weak deformation of the Bakkejord Diorite conflicts with a scenario of tight infolding. (3) A cover sequence that was deposited on the Gråtind Migmatite or Bakkejord Diorite and has acted as a thrust zone. We favour this interpretation of the Mjelde-Skorelvvatn Zone as a tectonic boundary because it best fits the preferred kinematic model outlined above.

On a wider scale, Binns (1983b, 1985) interpreted the metasupracrustal rocks of the Mjelde-Skorelvvatn Zone and Steinskardtind Unit (Fig. 1) on Kvaløya as Palaeoproterozoic formations, and correlated them with the Ringvassøya Greenstone Belt (Fig. 1). Metavolcanic members of the Ringvassøya Greenstone Belt have U-Pb deposition ages of 2848.5 ± 4 Ma and 2835 ± 14 Ma (Motuza et al. 2001), placing them in the Archaean. Zwaan & Walker (1996) correlate the Gråtind Migmatite on Kvaløya with the Dåfjord Gneiss on Ringvassøya, which has a crystallisation age of 2842 ± 2 Ma. The petrographic similarity of these two gneiss units (Zwaan 1989; Zwaan 1992a; Bergh & Armitage 1998) is at present the only support for such a correlation, while a kinematic kinship is apparent for the Mjelde-Skorelvvatn Zone and the Torsnes and Astridal Shear Zones on Senja (Nyheim et al. 1994; Zwaan & Bergh 1994; Pedersen 1997). The suggested geological link between Kvaløya, Senja and Ringvassøya implies a formerly extensive Archaean supracrustal sequence, including the Senja supracrustal belts, the Ringvassøya Greenstone Belt (Bergh & Armitage 1998) and the Skorelvvatn Formation, lying on crystalline basement with multiphase deformation of this basement-cover package during the Palaeoproterozoic tectonic event(s) deduced in our kinematic model. Further age determinations for the deposition of supracrustal belts on Kvaløya and Senja will greatly advance our understanding of their original setting.

Le Pichon et al. (1977) and Stearn & Piper (1984) argued that Fennoscandian basement north of present 65° latitude, possibly represented by northern Senja and Kvaløya, may have been thrust eastwards in Caledonian time to high structural levels. Dallmeyer (1992) supports this argument in his observation that Caledonian overprinting was minor at the crustal depths now represented by the surface rocks (Dallmeyer 1992). We observe, however, that major shear zones of the West Troms Basement Complex appear to be in line with aeromagnetically defined shear-lens patterns in Sweden and Finland, interpreted as expressions of intense Svecofennian transpressional strike-slip movement that created the Bothnian-Senja Fault Zone (Henkel 1987). This continuity of major shear zones through the West Troms Basement Complex, although obscured by the Caledonide nappe pile, suggests that the West Troms Basement Complex is autochthonous or has only been subjected to minor Caledonian displacement. Another possible cause of deep exhumation of the West Troms Basement Complex is footwall uplift associated with the Mesozoic Kvaløysletta-Straumsbukta Fault (e.g. Olesen et al. 1997). This structure dips east, juxtaposing the West Troms Basement Complex and the Caledonides (Fig. 1) with a minimum vertical throw of 2000 m, and was a master fault in the Palaeozoic unroofing of the orogen (Forslund 1988).

Conclusions

The structural architecture of the Mjelde-Skorelvvatn Zone on Kvaløya, Troms, is tentatively interpreted as the legacy of a Svecofennian transpressional event. This interpretation is supported in the main by the minimum age of 1768 ± 4 Ma for a granitoid pegmatite dyke evidenced to be synchronous with lateral shearing (D_3). Further support is provided by geometric and metamorphic arguments for a preferred model in which D_1, D_2 and D_3 are increments of a progressive partitioned event in amphibolite-facies conditions, involving WSW-ENE oblique contraction (transpression) across a NNW-SSE striking boundary represented by the Skorelvvatn Formation. The proposed kinematic model for the Mjelde-Skorelvvatn Zone is compatible with established models for the Senja Shear Belt, i.e. the Torsnes Shear Zone on Kvaløya (Zwaan 1995; Nyheim et al. 1994) and Astridal Shear Zone (Pedersen 1997) on Senja (Fig. 1). There are also similarities to a hypothetical model for the Ringvassøya Greenstone Belt (Zwaan 1989; Bergh & Armitage 1998). The shear structures in each zone have somewhat dissimilar configurations, but these can be explained by the slightly varied orientations of the metasupracrustal belts relative to the regional shear plane in the Palaeoproterozoic.

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