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Kinematic stratification in the hinterland of the central Scandinavian Caledonides

JANE A. GILOTTI

Geological Survey, New York State Museum, The State Education Department, Albany, NY 12230, U.S.A.

and

JOSEPH M. HULL

Department of Geography and Planning, SUNY-Albany, Albany, NY 12222, U.S.A.

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Abstract—A transect through west-central Norway illustrates the changing geometry and kinematics of collision in the hinterland of the central Scandinavian Caledonides. A depth section through the crust is exposed on Fosen Peninsula, comprising three tectonic units separated by two shear zones. The lowest unit, exposed in the Roan window, is a modestly deformed, Caledonian granulite complex framed by a subhorizontal décollement, with NW–SE oriented lineations and kinematic indicators showing top-to-the-northwest transport. The middle unit, the Vestranden gneiss complex, contains relict granulites, but was penetratively deformed at amphibolite facies to produce an orogen-parallel family of structures during translation on the décollement. Shallow plunging lineations on steep schistosities are subparallel to fold axes of the dominant, upright, non-cylindrical folds. A small component of sinistral strike slip is also recorded. In contrast, southernmost Fosen Peninsula contains an abundance of cover rocks infolded with Proterozoic basement in a fold nappe, with shallow, E-dipping schistosities, down-dip lineations, and orogen-oblique, top-to-the-west shear sense indicators. A NE-striking, sinistral shear zone separates the gneisses from southern Fosen. Deformation in the Scandian hinterland was partitioned both in space and time, with orogen-parallel extension and shear at middle structural levels and orogen-oblique transport at shallower levels.

INTRODUCTION

OROGEN-NORMAL transport dominates over longitudinal displacement in the thin-skinned, foreland portions of fold and thrust belts (Elliott 1976). Slip on subparallel thrust faults with a relatively constant direction and sense of slip produces a foreland-directed, bulk simple shear (plane strain) deformation on a regional scale (e.g. Mitra *et al.* 1988). The more hinterland portions of a collisional orogen, where penetrative, non-plane strains are more pronounced, show complex deformation patterns. Subhorizontal shortening may still dominate, but the principal extension direction commonly varies, and material may be transported along strike parallel to the orogen (e.g. Ramberg 1966, Bryant & Reed 1969, Dewey *et al.* 1986, Ellis & Watkinson 1987). Heterogeneous deformation on a crustal scale during continent–continent collision results in kinematic stratification within the orogenic belt. Compatibility must be maintained between the different structural levels within the framework of the collision-related boundary conditions.

The hinterland of the Scandinavian Caledonides, as exemplified by west-central Norway (Fig. 1), was constructed during the Scandian deformation phase during Siluro-Devonian time. In this paper, we discuss new results along a structural transect across Fosen Peninsula, from the cover-dominated region along Trondheimsfjord, to the high-grade basement rocks on the coast. Our study area traverses three different structural levels within the hinterland, and allows us to document the crustal stratification of deformational geometries and kinematics during the mid-Paleozoic Scandian orogeny.

In this paper, we describe the geology and structural fabric of five major tectonic elements along the traverse (Fig. 2). From north (deepest level) to south, these tectonic elements are: the Roan window; the Einarsdalen framing thrust; the Vestranden gneiss complex; the Møre-Trøndelag fault system; and the Inner Fosen basement–cover complex. From the transect data, we develop a three-dimensional model for the Scandian deformation in this part of the hinterland. Our studies document a significant component of strike-parallel extension for this area. Various models for orogen-parallel transport are discussed.

GEOLOGICAL SETTING

The Western Gneiss Region (Fig. 1) exposes the deepest structural level in the Scandinavian Caledonides. Sm–Nd and U–Pb geochronology, as well as thermobarometric studies from eclogites in the southern part of the Western Gneiss Region, attest to a widespread, high-pressure event during Scandian continent–continent collision (Griffin *et al.* 1985, Gebauer 1990). Although there are many petrologic and chronologic studies from the Western Gneiss Region, the internal structure of the gneisses is poorly known and the extent

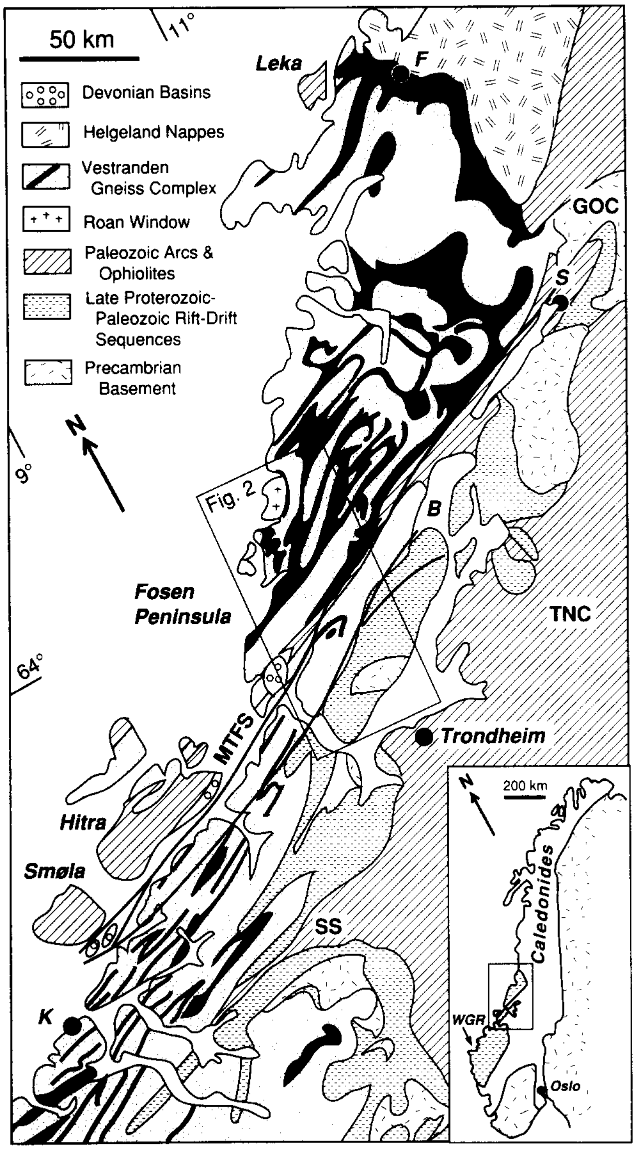

Fig. 1. Geologic map of the western portion of the central Scandinavian Caledonides, modified from Wolff (1976) and Solli (1990), with additional information from Bryhni (1989). Vestranden gneiss complex; orthogneisses (dots), paragneisses and amphibolites (black). Abbreviated names are: B = Beitstadfjord, F = Folderoid, GOC = Grong-Olden Culmination, MTFS = Møre-Trøndelag fault system, S = Snåsa, TNC = Trondheim nappe complex, SS = Surnadal synform, and K = Kristiansund.

of Scandian overprint on Proterozoic crust has been widely debated. Recent 1:50,000 scale mapping by the Norwegian Geological Survey's Nord-Trøndelag Project (Boyd 1986) has provided new structural data for the northern part of the Western Gneiss Region known as Vestranden (Fig. 1).

Our transect across the central Norwegian hinterland traverses the western part of Fosen Peninsula (Fig. 2) in Vestranden. Three out of six new 1:50,000 map sheets that cover the transect area have now been published (Fossen *et al.* 1988, Grønlie & Möller 1988, Thorsnes & Grønlie 1990), and the others are in preparation. Reconnaissance mapping prior to this project is summarized by Roberts (1986). Both authors have participated in the Nord-Trøndelag Project: Hull in the Leksvik area, south of the Verran fault; and Gilotti in the Vestranden gneisses, south of the Roan window (Fig. 2). In addition to mapping and structural analysis, we have collected detailed structural observations from key areas in the western part of Fosen. Despite the uncertainties in the timing of events, we feel that this particular section provides an excellent view of the geometry and kinematics in the hinterland of the Scandinavian Caledonides.

TECTONIC ELEMENTS

Roan window

The Roan window (Fig. 3) is the deepest structural level exposed in Vestranden. The area has been described in detail by Fossen *et al.* (1988), Grønlie & Möller (1988) and Möller (1988, 1990). The dominant lithologies are granitic, quartz monzonitic, quartz monzodioritic and charnockitic orthogneisses, some of which are Middle Proterozoic (1640–1650 Ma) in age (Möller 1990, Dallmeyer *et al.* 1992). These rocks were intruded by layered basic complexes during the Early Ordovician (Möller 1988). Modest strains have helped to preserve some intrusive relationships, but more detailed geochronology is needed to work out the exact intrusive sequence.

The Roan window is an elongate, NE–SW-trending dome characterized by a weakly developed, orogen-parallel family of structures (Möller 1988). The granulites tend to be massive, but a weak foliation defined by oriented clinopyroxene can be recognized locally. A semi-penetrative, amphibolite facies schistosity mimics the dome, with shallow orientations in the center steepening towards its perimeter (cf. fig. 12 in Möller 1988). This schistosity is axial planar to isoclinal folds of gneissosity, and increases in intensity towards the upper tectonic boundary. Minor hinge lines and amphibolite facies mineral lineations plunge shallowly towards the northeast and southwest in the core of the window (Fig. 3), but some are oriented NW–SE as the overlying contact is approached.

High-pressure granulite facies assemblages are well preserved within the Roan window, despite the widespread amphibolite facies overprint. A distinctive kyanite + clinopyroxene + garnet assemblage from the layered mafic complex at Kråkfjord (Fig. 3) gives peak metamorphic temperatures of $\approx$870°C and pressures of $\approx$1.4 GPa (Johansson & Möller 1986). The high-pressure metamorphism is interpreted to be Scandian in origin based on a Sm–Nd garnet–clinopyroxene–orthopyroxene whole-rock isochron of 432±6 Ma (Dallmeyer *et al.* 1992).

Retrogression of the kyanite-bearing high-pressure assemblage to sapphirine + corundum + orthopyroxene + plagioclase symplectites in the sillimanite stability field indicates that rocks within the Roan window underwent near isothermal uplift from approximately 1.4 to 0.8 GPa. The Roan window subsequently cooled through 500°C at about 400 Ma in the andalusite stability field (Möller 1990, Dallmeyer *et al.* 1992).

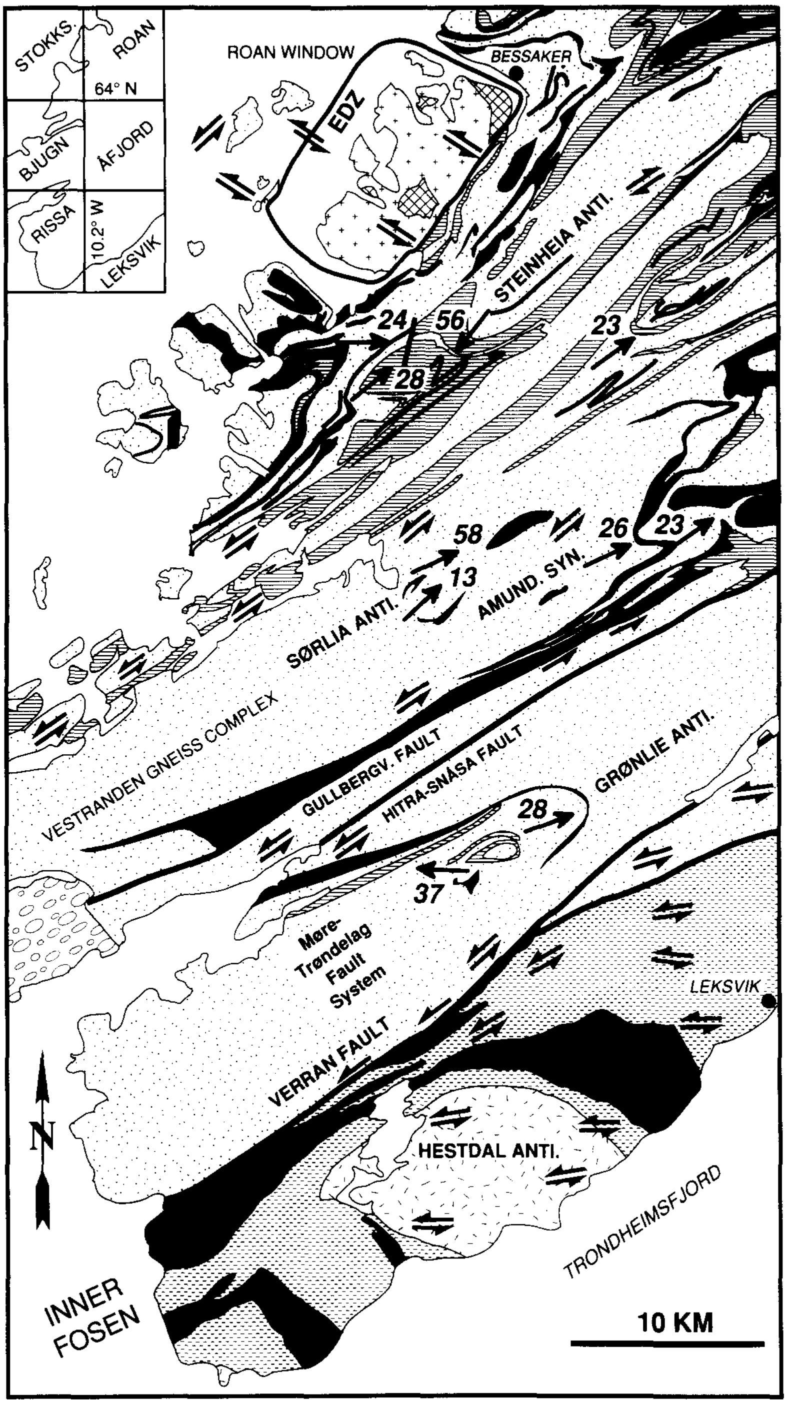

Fig. 2. Simplified geologic map and major structural elements of the transect area, based on recent 1:50,000 mapping by NGU (map sheets shown in inset). The Vestranden gneiss complex (VGC) includes orthogneisses (stippled), amphibolites (black) and paragneisses (striped). Plunges of fold axes for major upright folds are shown. Kinematic indicators are associated with steeply dipping fabrics in the VGC. The Inner Fosen complex includes basement orthogneisses (random ticks), metasediments (dashed), and an intrusive suite (black). The Devonian Fosen basin is also shown (ellipses). EDZ = Einarsdalen deformation zone.

Einarsdalen deformation zone

The Einarsdalen framing thrust is an approximately 1 km thick zone of intense strain that separates the rocks of the Roan window from the overlying paragneisses, orthogneisses and amphibolites of the Vestranden gneiss complex (Fig. 3). Granitoid orthogneisses comprise over half of the rocks in the Einarsdalen deformation zone. Marbles, calc-silicates, garnet–biotite schists, garnet–kyanite–plagioclase gneisses and amphibolites dominate the paragneisses. The base of the Vestranden gneiss complex at Einarsdalen is located at the lowermost paragneiss. This division is somewhat arbitrary because mylonitized granitic orthogneisses within the Einardsalen deformation zone could be derived from either the footwall or hangingwall (Fig. 3).

The Einarsdalen deformation zone is domed to form the Roan window (Fig. 4). Gneissosity, schistosity and mylonitic foliations are subparallel; these planar fabrics dip moderately towards the southeast on the southeast-

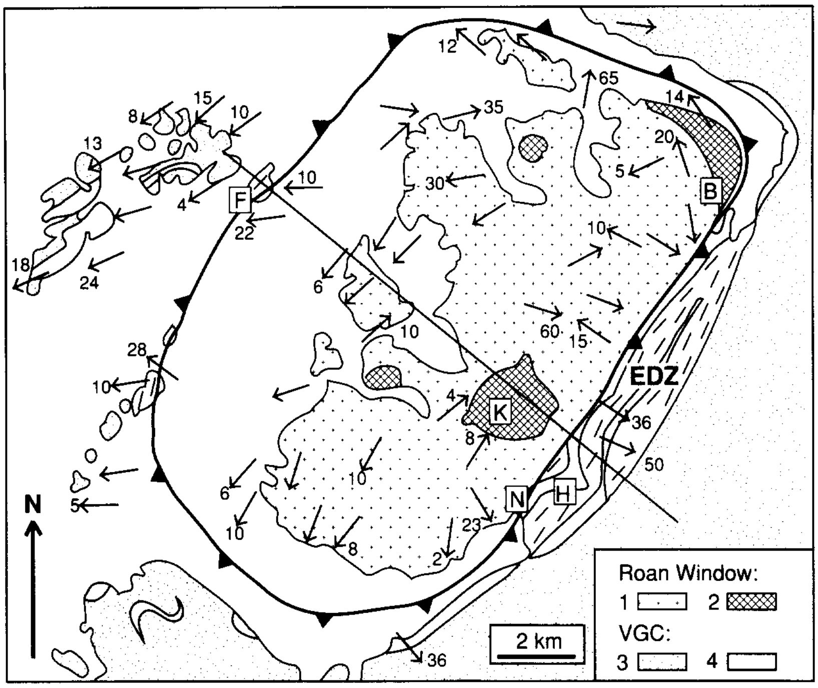

Fig. 3. Geologic sketch map of the Roan window (1, quartzofeldspathic orthogneiss; 2, basic intrusive complexes) and the overlying Vestranden gneiss complex (3, acid–basic orthogneiss; 4, supracrustal sequences). The Einarsdalen deformation zone (EDZ) contains highly tectonized orthogneisses (dashed) and two continuous supracrustal bands. The EDZ is interpreted as a thrust whose lower boundary is mapped at the lowest supracrustal unit. Arrows are mineral and stretching lineations. Locations mentioned in the text are: B = Brandsfjorden, F = Farmansöya, H = Hannafjellet, K = Kråkfjord, and N = Nordskjør. Modified from Möller (1988).

ern side of the window, and to the northwest on the islands (Figs. 4a & b). The Einarsdalen deformation zone is not exposed in the northeast and southwest, but foliations are assumed to parallel those in the overlying gneisses. Kyanite, sillimanite and hornblende form mineral lineations, while stretching lineations are defined by aggregates of plagioclase, K-feldspar, quartz and/or biotite, which appear as streaky rods lying on the foliation surfaces. Lineations fan 120° about the down-dip direction on the planar fabrics, with no systematic orientation according to lineation type (Fig. 4c). Deformation, gauged by grain size variation, is heterogeneous within this 1 km thick zone. Pin-stripe gneisses, with very fine plagioclase and quartz ribbons, are interpreted as high-grade, high-strain zones. Discrete zones of greenschist to amphibolite facies mylonites are common in the quartzofeldspathic gneisses, where they inosculate around less deformed lithons of orthogneiss. For

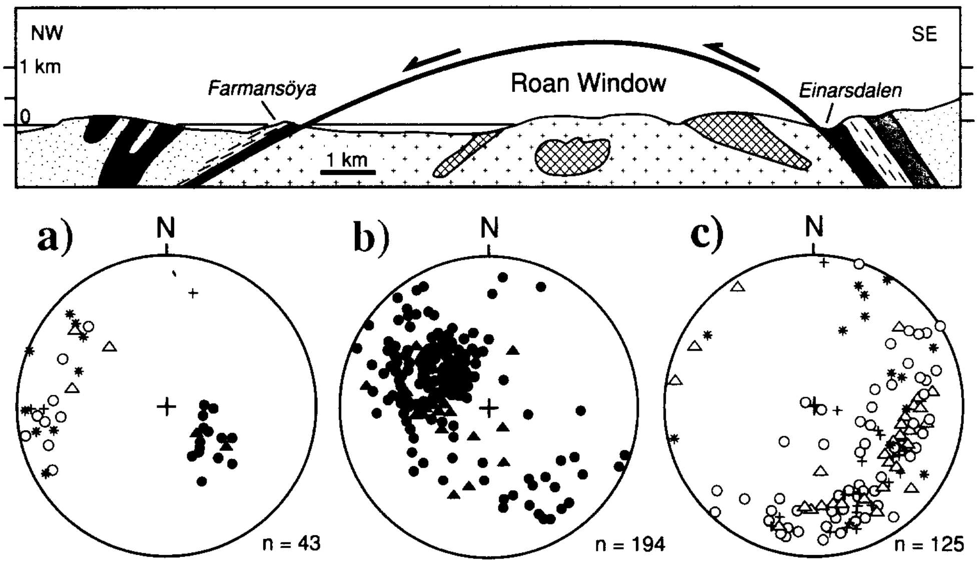

Fig. 4. Schematic cross-section through Fig. 3, modified from Möller (1988). (a) Structural elements for the northwest island exposures of the Einarsdalen deformation zone. (b) Foliations and (c) lineations for the southeast exposure of the Einarsdalen deformation zone, where: ● = gneissosity or schistosity, ▲ = mylonitic foliation, △ = mylonitic lineation, ○ = stretching lineation, + = kyanite lineation, and * = hornblende lineation. Note: Stereonets in all the figures are lower-hemisphere, equal-area projections.

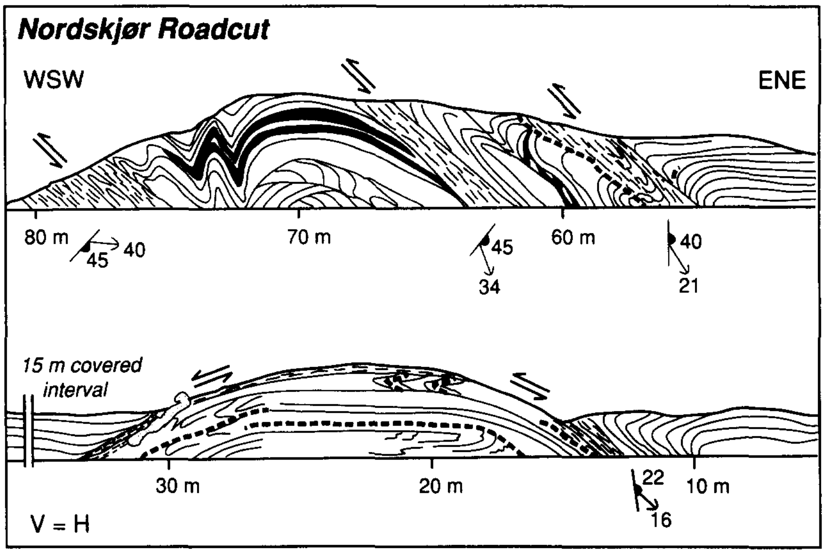

Fig. 5. Schematic diagram of an 80 m long roadcut at Nordskjør (locality indicated on Fig. 3). Strike and dip symbols correspond to the orientation of the mylonitic foliations and lineations measured from the individual shear zones (dashed pattern). The mylonite zone at 13 m is folded, but the other three are more typical, subparallel, planar zones. Pegmatite layers (heavy dashed lines) are folded (at 20 m) or localized along shear zone boundaries (at 14 and 57 m). A late pegmatite cuts a shear zone at 30 m. Amphibolite gneiss at 60 m (black) is folded together with gray orthogneisses (white and gray).

example, four such mylonite zones, 1–3 m thick, are exposed in an 80 m long road cut at Nordskjør (Fig. 5). The spread of lineations and individual mylonite zone orientations indicate that the Einarsdalen deformation zone is not a single zone of simple shear, but involves complex strain patterns, or numerous, smaller, simple shear zones with a variety of transport directions.

Kinematic indicators are abundant in the amphibolite facies tectonites which dominate the Einarsdalen deformation zone, and range from macroscopic to microscopic in scale. The vast majority show NW-directed transport of the hangingwall relative to the underlying Roan window. Discrete mylonite zones, where discordant gneissosity curls into parallelism with the mylonitic foliation, give the most reliable indication of shear sense (Fig. 5). The mylonites themselves contain numerous asymmetric delta and sigma feldspar augen, hornblende fish, and *S–C* fabrics. In addition, NW-verging fold trains are well exposed at Nordskjør (Fig. 5) and on a 200 m high cliff section at Hannafjellet (Fig. 6).

The Einarsdalen deformation zone was established at peak metamorphic conditions. The same sapphirine-bearing assemblage that was used to obtain peak *P–T* estimates in the Roan Window (Johansson & Möller 1986) is found in the Einarsdalen deformation zone at Brandsfjorden (Fig. 3) as an *L–S* tectonite with coarse kyanite blades defining the mineral lineation. Two lenses of mafic granulite from the paragneisses in the Einarsdalen deformation zone yield peak metamorphic conditions of $T \approx 860°C$ and $P \approx 1.4$ GPa (Möller 1990). Common garnet–biotite–quartz–K-feldspar–plagioclase–kyanite schists indicate similar peak conditions, with retrograde textures (e.g. biotite + sillimanite replacing garnet and sillimanite and andalusite after kyanite) indicating continued fabric-forming events at lower pressures and temperatures (Möller 1990). Shear sense indicators have not been found in the granulite facies tectonites, but the same orientations of fabric elements at all grades, coupled with the rapid uplift indicated by the *P–T–t* path, gives us no reason to suspect any transport other than top-to-the-northwest in the Einarsdalen deformation zone.

Vestranden gneiss complex

The Vestranden gneiss complex is a complicated tectonic unit that lies north of the Verran fault of the Møre-Trøndelag fault system and above the Roan window (Figs. 1 and 2). There are no detailed or comprehensive studies of the Vestranden gneiss complex in the transect region (Fig. 2), but some information can be found in Johansson (1986a,b), Möller (1988), Piasecki & Cliff (1988), Schouenborg (1989) and Thorsnes & Grønlie

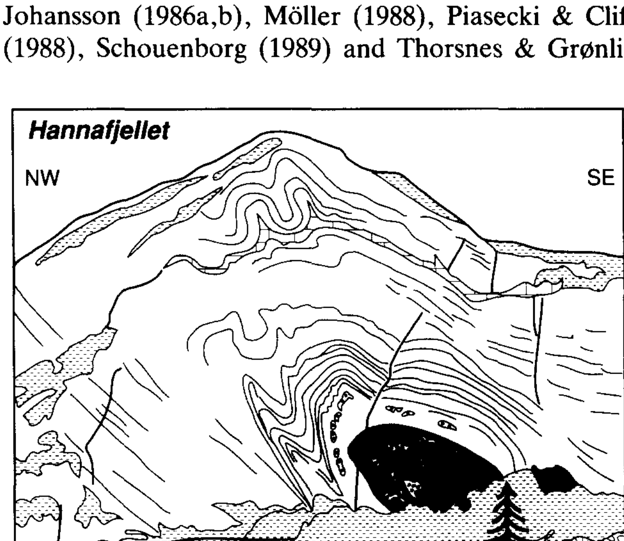

Fig. 6. Northwest-verging, disharmonic folds on the 200 m high cliff at Hannafjellet in the Einarsdalen deformation zone (locality indicated on Fig. 3). The rocks are gray orthogneisses with mafic lenses (black) and pegmatites (dotted). An orange marble unit (pattern) near the top of the cliff forms a décollement surface beneath two box-like, rabbit ear folds.

(1990). The three main lithologic units are orthogneisses, paragneisses and amphibolites. The orthogneisses are mostly grey migmatitic and non-migmatitic quartzofeldspathic gneisses with lesser amounts of red granitoid gneiss. The paragneisses are similar to those described from the Einarsdalen deformation zone. The amphibolites are quite variable, ranging from massive mafic units to banded intermediate and felsic rocks; their protoliths are unknown but some were probably dikes and sills. Intrusive relationships between the various gneisses have been obscured by deformation. Pods of relict, high-pressure, mafic granulites and coronitic metagabbros are scattered throughout the Vestranden gneiss complex (Fig. 7, and Johansson 1986b).

The map pattern of the Vestranden gneiss complex is dominated by late, tight to isoclinal, upright folds with both NE- and SW-plunging fold axes (Figs. 2, 7 and 8). These upright folds deform lithologic units, gneissosity and a schistosity that is subparallel to the gneissosity. An earlier fold phase is suggested by outcrop scale, isoclinal folds of gneissosity with hinge planes parallel to the foliation. Map scale examples of polyphase folding, such as those in the hinge zone of the Steinheia antiform (Fig. 7) can be found regionally, commonly forming Ramsay (1967, p. 530) type 3 interference patterns of refolded isoclines.

The plunge of individual upright folds varies considerably (Fig. 2). For example, the Sørlia antiform and the Steinheia antiform are steeply plunging folds, while the Amunddalsvatnet synform is an open, shallowly plunging fold. A large, doubly-plunging, tight, upright fold, the Grønlie antiform (Fig. 8), dominates the southern part of our transect. The variation in fold plunge is indicative of moderately non-cylindrical folds on the regional scale. Fold inclination changes from upright to shallowly inclined and recumbent near the Roan window; an example is the well-exposed fold at Bessaker (fig. 3 in Möller 1988).

In contrast to the rocks in the Roan window, the Vestranden gneisses have been thoroughly recrystallized and deformed under amphibolite facies conditions. The gneisses are *L–S* and *L* > *S* tectonites with a pervasive, orogen-parallel, amphibolite facies fabric. Gneissosity and schistosity are subparallel and generally steep with a NE–SW strike (Fig. 9) except where they are folded in the hinge zones of upright folds. New axial planar fabrics are rare in these hinge zones. Lineations lying on the foliation planes are shallowly plunging towards the northeast and southwest (Fig. 9). Hornblende mineral lineations are ubiquitous and subparallel to stretching lineations defined by ribbons and streaky aggregates of quartz and feldspar. The marked parallelism between mineral and stretching lineations and the hinge lines of major and minor folds (also noted by Ramberg 1966) defines an orogen-parallel structural family for the Vestranden gneiss complex.

A heterogeneous component of non-coaxial strain is indicated by widespread, though not pervasive, asymmetric structural elements. The non-coaxial structures include: shear bands; asymmetric boudinage; discrete black, garnet–biotite-rich mylonite zones in orthogneiss; and mylonitic paragneisses with mostly sigma-type augen. The mylonite zones are parallel to typical Vestranden fabrics, with steep, orogen-parallel foliations and subhorizontal lineations. The large majority of cases indicate sinistral, strike-slip displacements (Fig. 2).

There are no published radiometric dates from the Vestranden gneiss complex on our transect (Fig. 2), though Precambrian ages are suspected for much of the orthogneiss by comparison with other areas. In northern Vestranden near Folderoid, gray tonalites are at least 1820 Ma old (Schouenborg *et al.* 1991); a granite dike at Osen, 10 km north of the Roan window, is 1630 Ma old (Schouenborg *et al.* 1991); and the Geitfjellet granitic augen gneiss near Grong (Fig. 1) is 1829 Ma old

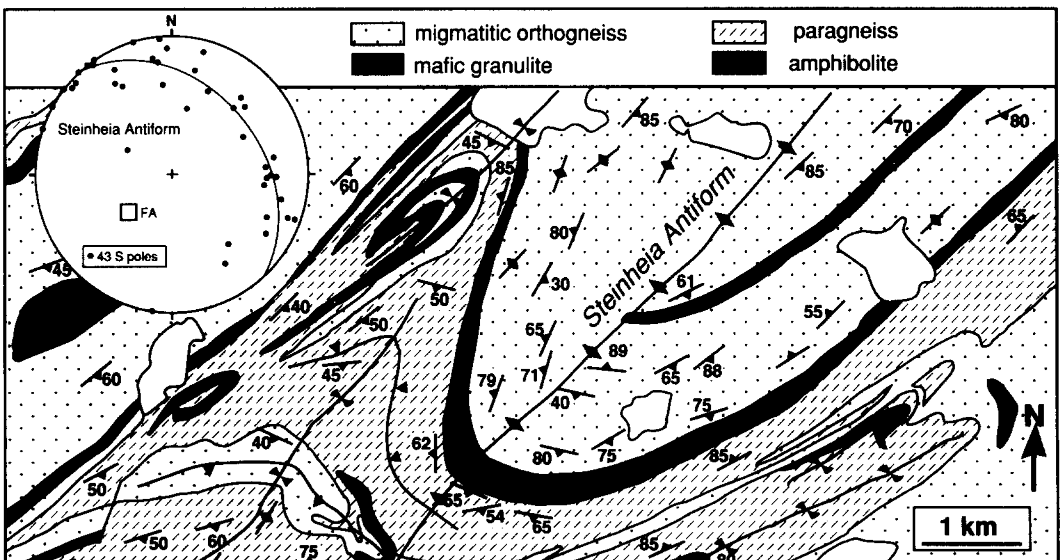

Fig. 7. Geologic map showing Ramsay type 3 refolded isoclines and a steep, southeasterly-plunging, upright antiform at Steinheia in the Vestranden gneiss complex. The area is located on the Roan 1:50,000 sheet (Fossen *et al.* 1988).

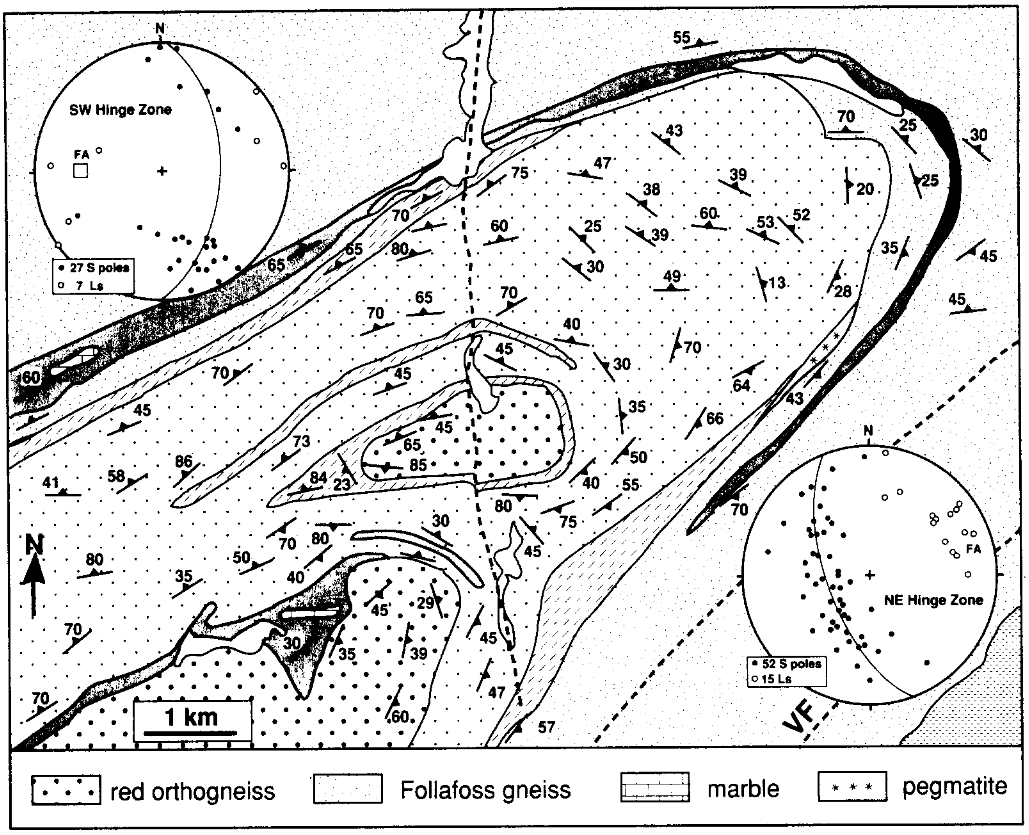

Fig. 8. The Grønlie antiform in the Vestranden gneiss complex is a doubly-plunging antiform. The area is located in the southeast corner of the Åfjord 1:50,000 sheet (Thorsnes & Grønlie 1990). VF is the Verran fault, and the dashed lines are other brittle faults. Other symbols are the same as in Fig. 7.

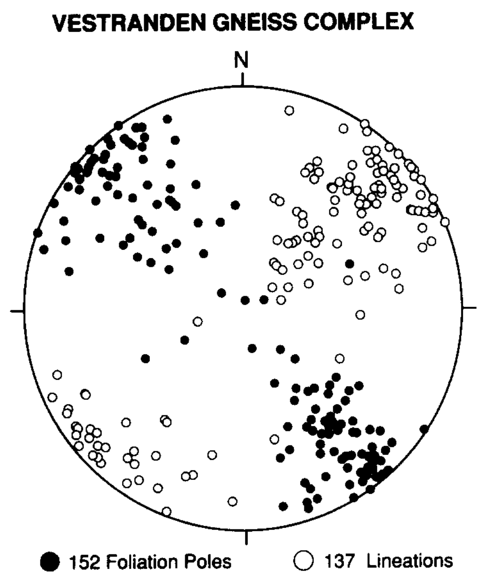

Fig. 9. Stereonet showing schistosities and lineations from fold limbs in the Vestranden gneiss complex in the region covered by Fig. 2.

(Johannson 1986a). An Ordovician age (460 ± 4 Ma) has recently been obtained (Thorsnes *et al.* in review) from the Follafoss tonalite at a locality along strike with the Grønlie antiform (Fig. 8); a diorite gneiss near the Devonian basin on Fosen Peninsula is also Ordovician (P. Bickford personal communication, see also Tucker 1988). Undeformed pegmatites cutting the folded gneisses have been dated at two localities north of the Roan window yielding ages of 401 ± 3 Ma (Schouenborg 1988) and 398 ± 3 Ma (Dallmeyer *et al.* 1992).

The timing of deformation is poorly constrained; but if the granulite facies assemblages found in scattered mafic pods are of the same Scandian age as those in the Roan window (Dallmeyer *et al.* 1992), then the amphibolite facies metamorphism and deformation must be Scandian as well. We know that the upright folding is Scandian because the Ordovician Follafoss gneiss is folded (Fig. 8), but a Precambrian age for an early stage of isoclinal folding and fabric formation cannot be ruled out. An upper limit for the large-scale folding is given by the pegmatites, and corroborated by $^{40}Ar/^{39}Ar$ hornblende ages ranging between 413 ± 3 and 386 ± 6 Ma (Dallmeyer *et al.* 1992) which indicate cooling of the Vestranden gneisses through 500°C at approximately the same time. Muscovite and biotite in a mylonitic shear zone, and muscovite in a deformed pegmatite, both from the Bjugn map sheet, yielded single crystal, Rb–Sr model ages of approximately 390–365 Ma (Piasecki & Cliff 1988).

Møre-Trøndelag fault system

The Møre-Trøndelag fault system (Figs. 1 and 2) is considered by many authors to be a first-order tectonic boundary separating Vestranden gneisses to the north from predominantly supracrustal rocks to the south (e.g. Grønlie & Roberts 1989). The Møre-Trøndelag fault system consists of three major faults, the Hitra-Snåsa, Gullbergvatn and Verran faults, with subsidiary splays and imbricates. Despite the importance of the Møre-Trøndelag fault system in regional tectonic models (e.g. Grønlie & Roberts 1989, Andersen *et al.* 1991, Séranne 1992), there have been few studies of its tectonites, minor structures or kinematics (see Grønlie *et al.* 1991).

The Hitra-Snåsa fault (Figs. 1 and 2) is a 300 km long, topographic lineament drawn from the Grong-Olden culmination to Smøla in the southwest. To the northeast, the fault separates cover rocks from gneisses; however, on our transect (Fig. 2), the same Ordovician orthogneisses are found on either side. We have examined several localities along the Hitra-Snåsa fault. At the lineament trace, gray orthogneisses are partially retrogressed to chlorite- and epidote-bearing assemblages, and gneisses are cut by a network of thin, steeply dipping fractures. Gneissosity maintains a relatively constant orientation and is not disrupted in this 100 m wide zone. Estimates of the magnitude of slip on this zone of distributed faults are difficult to make without offset markers, but the modest tectonism observed in outcrop suggests relatively minor displacement along the Hitra-Snåsa fault in this area.

In contrast, the Gullbergvatn fault, located 5 km north of the Hitra-Snåsa lineament at a basement–cover contact, shows extensive tectonism. Northeast of Gullbergvatn, amphibolites, meta-agglomerates, metasandstones and minor semipelites are retrogressed to greenschists, chloritic–epidote phyllonites and button schists, while granodioritic gneisses are altered to phyllonites and quasiplastic mylonites. The zone of retrogression along this low-grade fault is up to several hundred meters wide, and individual mylonite zones are 1–10 m thick. Along Highway 715, the gneisses are progressively transformed over a distance of 200 m to protomylonites and pin-stripe mylonites as the fault is approached. The contact itself is covered, but chlorite schists, and many small chlorite and epidote rich faults and slickensides, deform the cover sequence. The dip of the mylonitic schistosity is towards the northwest, and varies from shallow to steep along the fault zone; lineations are subhorizontal. Shear bands are well developed in button schists and phyllonites, and indicate sinistral slip in map view (Fig. 2). The Gullbergvatn fault is shown as a thrust on the Trondheim and Åfjord map sheets (Wolff 1976, Thorsnes & Grønlie 1990), but we have seen no evidence for an earlier phase of thrusting along this contact.

The Verran fault is a topographic lineament extending 200 km from Beitstadfjord in the northeast to south of Smøla (Fig. 1). Cataclasites from the Verran fault have been described by Grønlie & Roberts (1989) and Grønlie *et al.* (1991). Higher grade or 'ductile' shear along the Verran lineament is at least as extensive as brittle faulting, however the geometry and kinematics of this component have received little attention (see Grønlie *et al.* 1991).

The Verran fault is spatially associated with a profound change in both structural geometry and tectonic units. South of the Verran fault, amphibolite facies schistosities dip shallowly eastward, with lineations

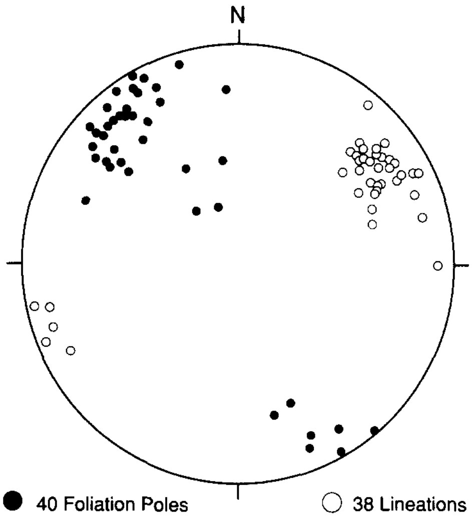

Fig. 10. Schistosity and mineral lineations in the zone of sinistral strike-slip along the Verran valley. Data taken from an approximately 2 km wide band along the valley.

plunging approximately down the dip (Hull & Adamy 1989, and next section). As the Verran valley is approached from the south, schistosity steepens and rotates into a typical NE strike, and mineral lineations plunge shallowly off to the northeast (eigenvector plunging 19° towards 060°) on steeply dipping schistosities (Fig. 10). The transition from more orogen-oblique, rather flat-lying structures to orogen-parallel, upright fabrics is approximately 3–6 km wide on our traverse.

The rotation of schistosity and lithologic contacts in the transition zone south of the Verran fault is consistent with sinistral shear (Fig. 2). Long, thick zones of orthomylonite are uncommon along the Verran valley, but many of the granitic rocks are protomylonites and augen mylonites. Kinematic indicators are particularly well developed in the discontinuous lenses and layers of coarse-grained, K-feldspar rich, granitic metaporphyries. Shear bands or extensional shear zones (mostly *C* surfaces) and asymmetric feldspar augen are the most common kinematic criteria. All of the outcrop-scale kinematic indicators show sinistral slip in map view (Fig. 2), consistent with the map-scale rotation of lithologic units and schistosity. Further evidence for sinistral strike slip can be found in Grønlie *et al.* (1991) and Séranne (1992).

There is no direct information on the ages of sinistral strike-slip components on the shear zones of the Møre-Trøndelag system. Stratigraphic, paleomagnetic and geochronologic evidence for post-Scandian brittle deformation along the Verran fault is summarized in Grønlie *et al.* (1991).

Inner Fosen basement–cover complex

The highest crustal level of the hinterland in our study area is exposed in the southern part of Fosen Peninsula or 'Inner Fosen' (Figs. 1 and 2), where supracrustal cover dominates over basement gneisses (Hull &

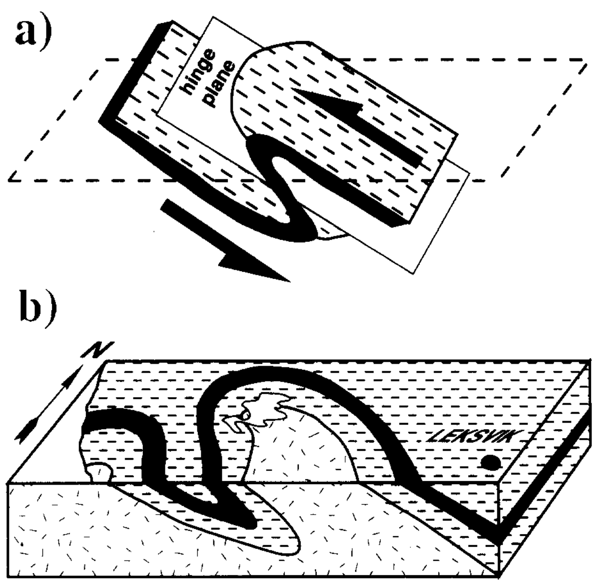

Fig. 11. Sketch indicating the development of Inner Fosen geology via a shallowly inclined, W-verging, non-cylindrical fold nappe (from Hull & Adamy 1989). Compare (b) with Fig. 2.

Adamy 1989). Proterozoic basement is composed of porphyritic to equigranular metagranitoids cut by several generations of granitic to basic dikes. Though large tracts of Late Proterozoic quartzite and arkose are shown infolded with orthogneisses on most compilation maps (e.g Roberts & Wolff 1981, Gee *et al.* 1985), these regions are actually mica-rich, equigranular orthogneisses. The cover suite is formed of metamorphosed calcareous sandstones and siltstones (often sulphide bearing), graphitic mudstone, limestone, quartzite and distinctive garnet–pyrrhotite–quartz beds. These metasedimentary rocks are superficially similar to the Gula Group of Trøndelag, however their ages and paleogeographic settings are unknown.

The cover has been intruded by a bimodal (?) suite of fine- to medium-grained gabbroic rocks and garnet-bearing, equigranular granodiorites. Cross-cutting relationships are locally well preserved and an intrusive sequence has been established (Hull & Adamy 1989). This intrusive complex may represent pre-Scandian, subduction zone plutonism comparable to the Ordovician arc on Smøla and Hitra (Gautneb 1988, Gautneb & Roberts 1989) and/or to Ordovician plutons in the Vestranden gneiss complex (Tucker 1988, Thorsnes *et al.* in review).

The Proterozoic orthogneisses are exposed in the core of a W-closing, shallowly inclined, tight to isoclinal antiform that also involves the supracrustal cover and intrusive rocks (Figs. 2 and 11). The antiform is quite non-cylindrical (Hull & Adamy 1989); the sheath-like morphology, geometry of minor structures, and kinematic framework (see below) are consistent with a basement–cover fold nappe formed in a shallow east-dipping, crustal scale shear zone.

The geometries of minor structures in the Leksvik map sheet are shown in Fig. 12. Schistosities dip subhorizontally to moderately north to east (Fig. 12a). Poles to schistosity show a diffuse, elongate maximum corresponding to a partial great circle girdle about an axis plunging shallowly east (10° towards 075°).

Mineral (mostly hornblende and feldspar) and stretching or shape lineations are subhorizontal and trend east-northeast (Fig. 12b), somewhat oblique to lineations within the Vestranden gneisses. In the Leksvik map sheet of Inner Fosen, the strong point maximum of lineations (11° towards 074°) is parallel to the axis of girdled schistosity. Mineral lineations are partially girdled within typical, N-dipping schistosities.

Four types of minor folds were discriminated in the field: (1) rare, unmodified folds of bedding; (2) intrafolial isoclines of bedding and/or gneissosity; (3) open to tight buckle folds of schistosity and gneissosity; and (4) rare folds of second schistosity (usually a crenulation cleavage). There are some small differences between the geometries of these four fold types. Most hinge lines of bedding folds and intrafolial isoclines trend approximately N–S (Fig. 12d), however the significance of this geometry is difficult to evaluate quantitatively, because of the relative paucity of data.

Hinge lines of buckle folds of schistosity show a similar distribution as mineral lineations (Fig. 12d); the two are almost always parallel in the field, and can be difficult to discriminate. These buckle folds plunge shallowly towards the east (eigenvector plunging 06° towards 085°), but are strongly girdled in the horizontal plane. The geometry of hinge planes of buckle folds mimics that of cleavage (Fig. 12c); the typical hinge plane dips shallowly towards the northeast (24° towards 030°). Poles to hinge planes are diffusely girdled about a shallow, E-plunging axis (18° towards 074°).

Planar zones of more intense deformation and mylonitization, dipping shallowly to moderately to the east and north, are widespread throughout the Leksvik map sheet, particularly along lithologic contacts. Deformation zones within all three map units have shallowly E-plunging lineations oblique to the S- to SE-trending transport directions shown in Hossack (1983) and Hossack & Cooper (1986). Kinematic indicators are best developed in the supracrustal sequence, but can be found in all three units. Asymmetric boudinage of amphibolitic sills, quartzite beds, quartz veins and other rigid bodies; *S–C* fabrics and discrete, extensional shear bands; sigma-shaped feldspar augen in metagranitoids; and asymmetric minor isoclines are the most abundant kinematic sensors. The kinematic criteria overwhelmingly show top-to-the-west transport (Fig. 2), consistent with the map-scale geometry. Similar geometries and kinematics have been described by Piasecki & Cliff (1988) from the northern part of Inner Fosen.

The basement orthogneisses in the core of the Hestdal antiform have been dated at approximately 1650 Ma (Tucker *et al.* 1990), however the ages of the other map units are unknown. All units have been penetratively deformed and metamorphosed in the amphibolite facies; biotite, garnet and kyanite are the index minerals in pelitic schists. Deformation and metamorphism are thought to be Scandian, as all rock units are involved,

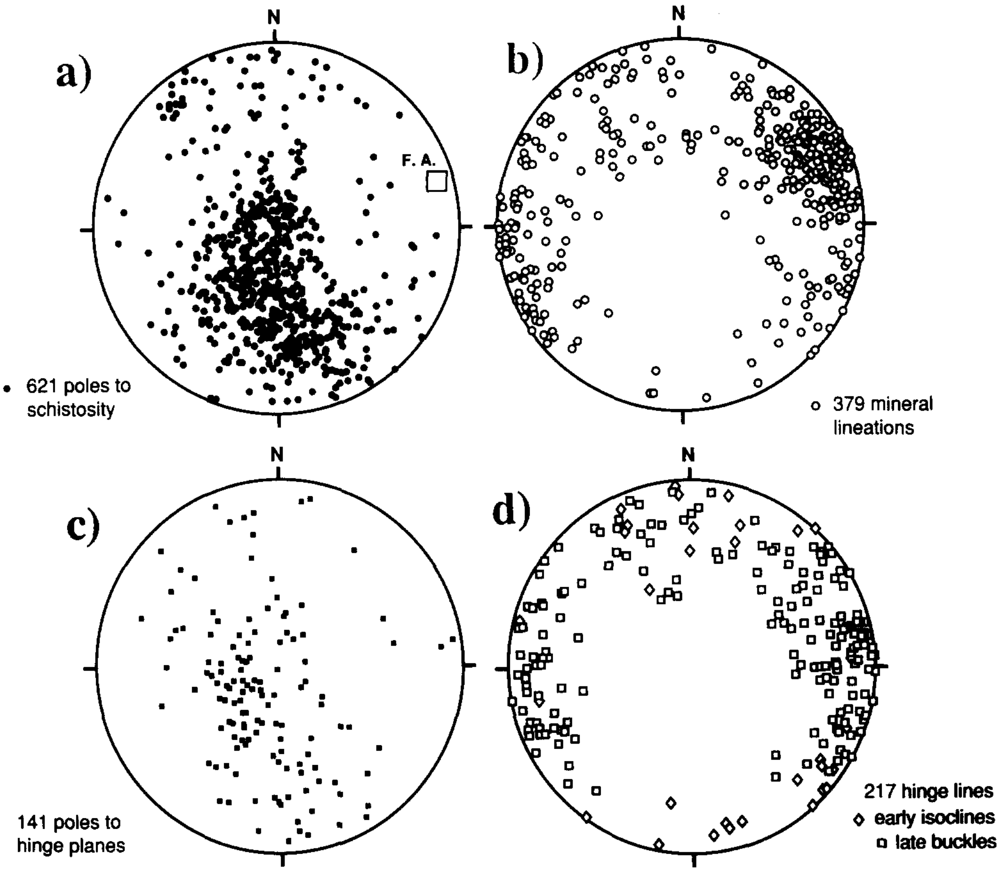

Fig. 12. Schistosities (a) and mineral lineations (b), and hinge planes (c) and hinge lines (d) of minor folds for the Leksvik map sheet within Inner Fosen. The large square is the calculated 'fold axis' for girdled schistosity. Hinge lines are subdivided into intrafolial isoclines of bedding and gneissosity (diamonds) and modified buckle-type folds of schistosity (squares).

but there is an obvious need for a comprehensive geochronologic study. Piasecki & Cliff (1988) reported a Rb–Sr model age of approximately 425 Ma for a muscovite book from a deformed pegmatite, sampled in the northern part of the Leksvik map sheet. This model age was interpreted by Piasecki & Cliff to date formation of the pegmatite during regional Scandian deformation.

STRUCTURAL INTERPRETATION

Each of the three major tectonic units across this part of Fosen Peninsula (Roan window, Vestranden gneiss complex, Inner Fosen basement–cover complex) has a different structural geometry and kinematic history (Fig. 13). These units are separated by major shear zones (Einarsdalen and Verran deformation zones). The geometry and kinematics of each of these regions can be interpreted in different ways, and we discuss our interpretations of each of them below.

Roan window and Einarsdalen deformation zone

The Einarsdalen deformation zone can be explained as a thrust or a low-angle normal fault. This zone was active under granulite facies conditions (Möller 1988), dated in the Roan window at 432 ± 6 (Dallmeyer *et al.* 1992), relatively early in the Scandian history (Roberts

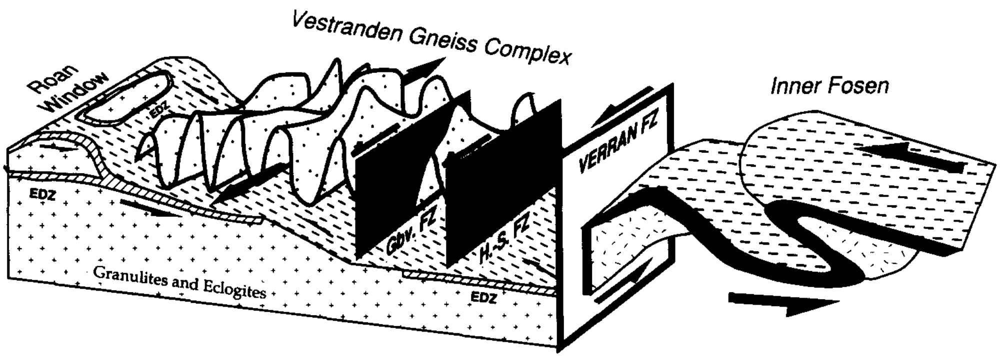

Fig. 13. Schematic diagram illustrating the main structural style and principal structural elements on the transect from Roan in the north to Inner Fosen in the south. The Einarsdalen décollement underlies the Vestranden gneiss complex; its branching relationship with the Verran fault is speculative.

& Wolff 1981, Gee & Sturt 1985, Hossack & Cooper 1986). This same high-grade metamorphism is locally preserved in the Vestranden gneiss complex, suggesting thrust repetition of fairly deep-level crust. Uplift and cooling of Vestranden, presumably related in part to extensional detachments, culminated in the Early to Middle Devonian (Hossack 1984, Steel *et al.* 1985, Dallmeyer *et al.* 1992). Thus early movement on the Einarsdalen deformation zone is perhaps 40 Ma older than extension on known detachments such as the Sunnfjord fault zone (Séranne & Séguret 1987).

We follow Möller (1988) in interpreting the Einarsdalen deformation zone as a thrust décollement, but suggest that the zone extends beneath the Vestranden gneiss complex. A geometrically analogous structure is the Alleghanian décollement that floors the Piedmont of the southern Appalachians (e.g. Harris & Bayer 1979, Cook *et al.* 1981, Iverson & Smithson 1982). The Alleghanian décollement is thought to glide in para-autochthonous, Cambro-Ordovician shelf sequences, deposited on a footwall of Grenville basement and Late Proterozoic rift facies rocks. The Einarsdalen décollement is located in a broadly similar metasedimentary sequence (marbles, pelites, etc.), however these metasediments are probably derived from the hangingwall, as similar supracrustal rocks are found throughout the Vestranden gneiss complex.

Several hypotheses can explain the topology of the Roan dome. Folding of the Einarsdalen décollement in a non-cylindrical, dome and basin fashion (Fig. 14a) might represent a late stage of the regional shortening seen in Vestranden. Folding of the décollement on a regional scale is supported by the local folding of individual mylonite zones (Fig. 5). Alternatively, the framing thrust is folded into a flat-topped antiform above a frontal ramp (Fig. 14b). A later thrust below the Einarsdalen décollement would step up and branch with the décollement; further slip would produce an insertion thrust (Geiser 1988) with a ramp antiform if cut-off angles were maintained. The narrow width of the Roan culmination suggests that the ramp is directly beneath the culmination, in this model. Lastly, the culmination may reflect the geometry and size of a footwall horse, and the footwall ramp may be located far to the south-east (Fig. 14c).

We favor some kind of imbrication model for the culmination, but recognize that all scenarios are difficult to evaluate. Models that invoke thrust imbrication require lateral ramps both to the northeast and southwest of Roan. If cut-off angles are maintained during displacement, the Einarsdalen décollement would mirror the geometry of the Roan imbricate or horse, and would show predominantly strike slip along NE- and SW-dipping segments. Unfortunately, these portions of the Einarsdalen décollement are not exposed (Fig. 3). A feature common to all the models is the relatively late timing of this culmination, as the shear sense and transport direction are constant (despite doming), and as the framing thrust was partially active under greenschist facies conditions.

The initial work of Möller (1988, 1990) and our detailed follow-up presented here documents displacement on a major décollement, but with an opposite transport sense to that assumed in previous tectonic models for this general area (e.g. Roberts & Wolff 1981, Gee & Sturt 1985, Hossack & Cooper 1986). Northwest-directed displacement on the Einarsdalen deformation zone is opposite to the SE-directed transport typically assumed for this region. As postulated by Hull & Adamy (1989), this 'backthrusting' may be part of a local pop-up related to subhorizontal shortening of the hinterland basement, as seen in other collisional orogens (e.g.

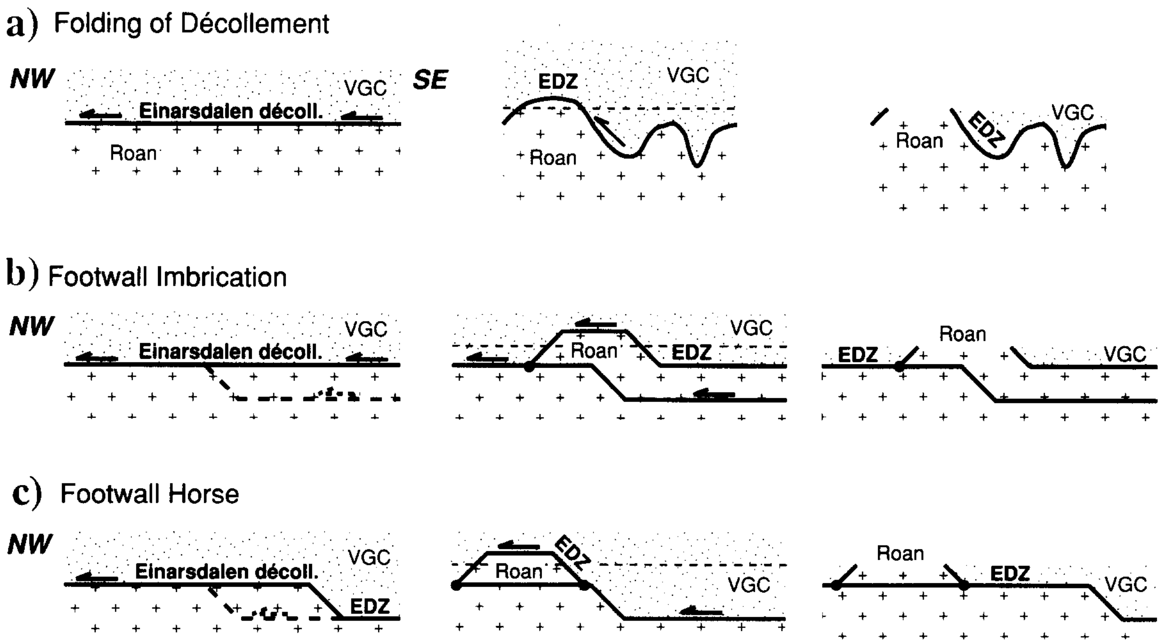

Fig. 14. Three alternatives for the origin of the Roan culmination. (a) Folding of the Einarsdalen décollement following thrusting. (b) Imbrication in the footwall of the EDZ, and development of an insertion thrust. (c) Development of a footwall horse, which may be far-travelled relative to the footwall step.

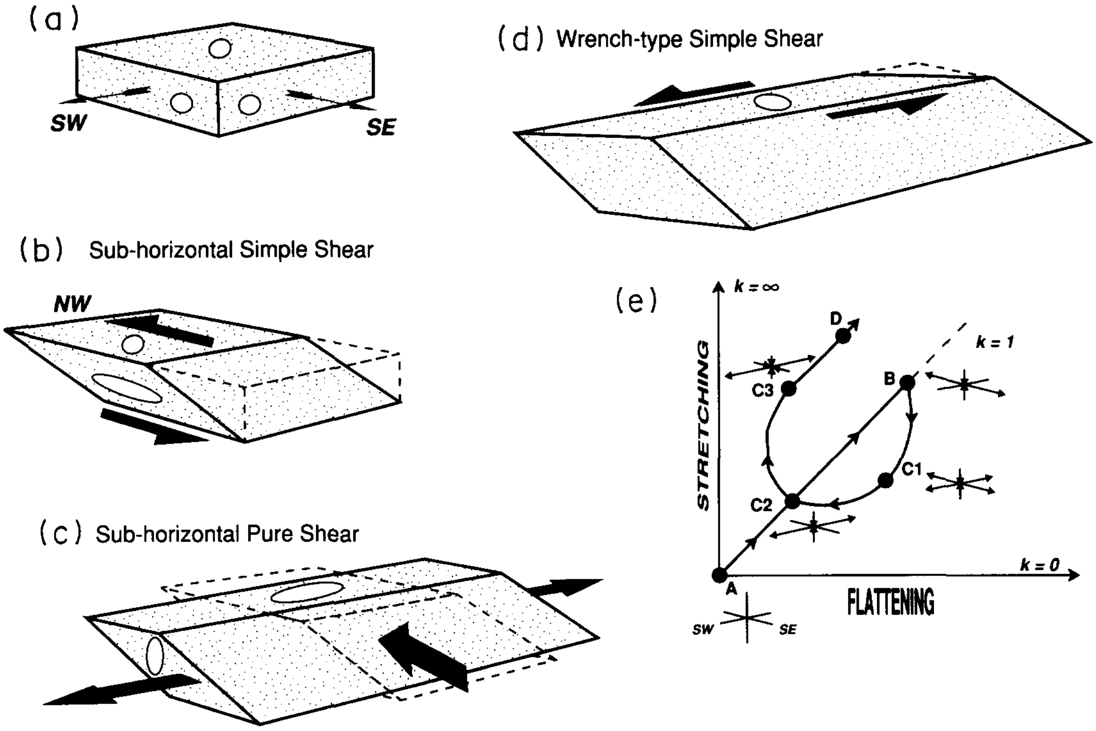

Fig. 15. Idealized progressive deformation of the Vestranden gneiss complex. (a) Initial stage. (b) NW-directed, subhorizontal shear. (c) Pure shear with NW shortening and NE–SW extension. (d) SW-directed shear on subvertical planes (wrench). (e) Schematic Flinn diagram illustrating one possible deformation path (among many). The geometry and style of finite strains at each stage are shown.

Coward *et al.* 1986). Alternatively, the Einarsdalen thrust may reflect a fundamental change in the transport symmetry in the western part of the orogen, a hypothesis we discuss at the end of this paper.

Vestranden gneiss complex

Early fabric forming and deforming events, producing recumbent isoclinal folds of gneissosity (e.g. Fig. 7) (fig. 3 in Möller 1988), pre-date the upright folding stage in the Vestranden gneisses. The ages of these older events are unknown; Early Proterozoic granitic rocks are present elsewhere within the Vestranden gneiss complex, and Early Proterozoic tectonism (e.g. Svecofennian) probably affected this area (see Schouenborg *et al.* 1991). However, given the growing evidence for widespread Paleozoic plutonism and very high grade Scandian metamorphism on our transect, we postulate that many of these early structures are Scandian. We further suggest that both the early formation and recumbent folding of gneissosity are related to progressive subhorizontal shear, directed towards the northwest (Fig. 15). We correlate this early component of NW-directed shear with the high-grade fabrics preserved within the Einarsdalen deformation zone.

One of the most important geometric features within the Vestranden gneiss complex is the parallelism between mineral or stretching lineations and hinge lines of the upright folds. There are a number of mechanisms that can produce this geometry. During simple shear, hinge lines of initially cylindrical folds are rotated into parallelism with the stretching direction to produce highly non-cylindrical sheath or tubular folds (e.g. Cobbold & Quinquis 1980). As a result, hinge lines fan around the stretching direction (i.e. the sheath axis). Upright folds within the Vestranden gneiss complex are somewhat non-cylindrical; however, the mineral lineations are curvilinear as well, tracking the hinge lines around the folds (e.g. Fig. 8), incompatible with the simple shear model.

In pure extension, such as found in the necks of salt domes, hinge lines of folds are parallel to the stretching direction, which is also a transport lineation. While *L*-tectonites are found within the Vestranden gneiss complex, most rocks exhibit both a well developed schistosity and mineral lineation. Folds in constrictional deformation are strongly polyclinal, with a wide variety of hinge plane geometries, whereas hinge planes of second folds on our traverse are uniformly upright over large regions.

Our model for the second, major deformational component in the Vestranden gneiss complex (Fig. 15c) is inspired by the plane strain, multilayer buckling experiments of Watkinson (1975). Subhorizontal shortening and extension between two fixed décollements results in folds whose hinge lines are subparallel to the extension direction (Fig. 16). The folds are rather harmonic away from the décollement, with lower amplitude to wavelength ratios than buckle folds in typical plane strain experiments without extension parallel to fold axes (Watkinson 1975). The fold trains are disharmonic with very low amplitudes near the décollement, and line lengths in fold profile suggest decreasing shortening towards the décollement. This topology requires an accommodation mechanism to balance the bulk subhorizontal shortening; either increased layer-parallel shortening towards the décollement by layer thickening, abundant contraction faults, or intense cleavage development, and/or more subhorizontal extension in layers adjacent to the décollement.

In Watkinson's experiments, the fixed detachments forced extension parallel to fold hinges at a high angle to the displacement or shear directions. In our model, the lower detachment is the Einarsdalen décollement (Fig. 16). Strike-parallel extension in the overlying Vestranden gneisses is perpendicular to the displacement direction on the Einarsdalen thrust. Strike-parallel extension above a basal décollement is also characteristic of large parts of the south-central Appalachian Piedmont (e.g. Secor *et al.* 1986, Hatcher & Tabor 1989). The upper décollement in Vestranden is now eroded. Candidates for this other detachment include an imbricate of the Vestranden gneiss complex; a thrust placing Paleozoic arcs and/or the Inner Fosen complex on top of the Vestranden gneisses; or a thrust placing the Trøndelag arcs and ophiolites onto both Inner Fosen and Vestranden.

The superposition of an early NW-directed simple shear and a subsequent NE–SW pure shear is illustrated on a Flinn diagram in Fig. 15. While there is a large array of deformation paths, depending upon the precise magnitudes and geometries of each of the strain components, we have chosen one path to illustrate the likely evolution of strain style and geometry in the Vestranden gneisses. Superposition of a horizontal pure shear initially results in a flattening finite strain, followed by plane strain and then constriction, producing a clockwise path on the Flinn diagram. The resulting finite strain state ($k > 1$) is consistent with the dominant $L > S$ fabric style seen in the Vestranden gneiss complex on our traverse.

The third deformation component within the gneisses of Vestranden is a SW-directed simple shear on subvertical planes; i.e. a SW-directed wrench. This wrench is usually localized along lithologic contacts within relatively narrow deformation zones. Though widespread in our study area, the overall magnitude of this simple shear is probably quite small. We do not expect substantial modification of the deformation path by this wrench (Fig. 15).

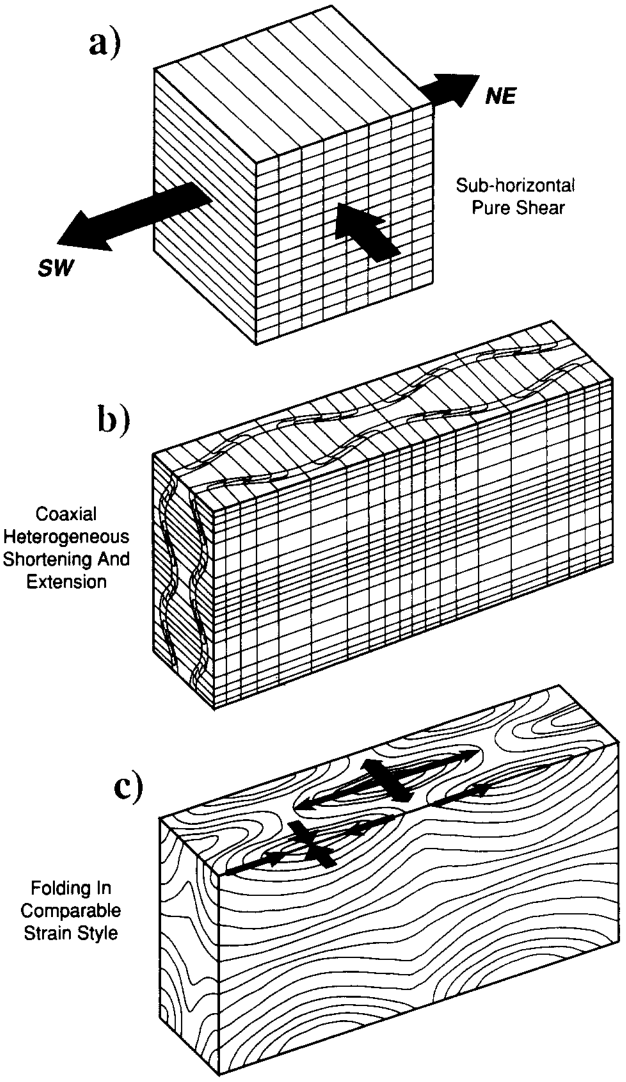

Fig. 17. Model of deformation zones and folds developed in heterogeneous shortening and extension, based in part on Bell (1981). Bulk coaxial deformation produces stretching faults of both shear senses.

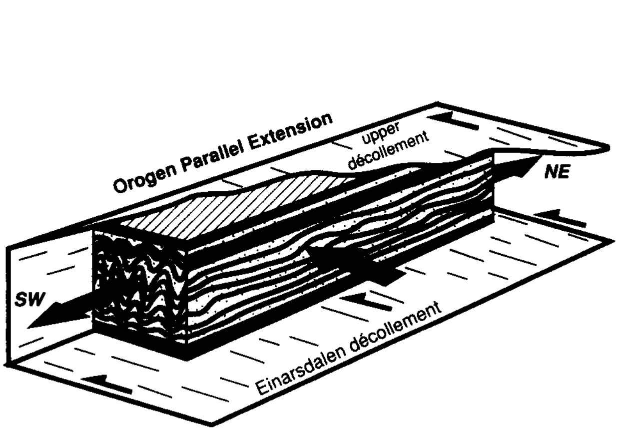

Fig. 16. Model for orogen-parallel (NW–SE) extension accompanying NW-directed shortening (bulk pure shear) between two fixed detachments, based on the experiments of Watkinson (1975). Extension is parallel to the hinge lines of folds that die out near the décollement.

Thus there are two components of transport parallel to orogenic strike; subhorizontal pure shear and a lesser component of sinistral strike-slip. Though strike-slip probably outlasted pure shear (see next section), the two components may have acted simultaneously; a transpressional deformation, overwhelmingly dominated by pure shear. The orogen-parallel wrench zones on our transect through the Vestranden gneiss complex could be the product of the same sub-horizontal pure shear that gave rise to axial extension (Fig. 17). In heterogeneous deformation, high strain zones of predominantly simple shear will enclose less deformed phacoids or augen exhibiting pure shear (Bell 1981). These shear zones are stretching faults (Means 1989), i.e. geometrically necessary deformation zones to maintain compatibility between blocks with different magnitudes and/or orientations of finite strain. As the wrench zones in our study area show consistent sinistral displacement, we expect the bulk shortening and extension to be somewhat non-coaxial, though very close to a bulk, coaxial pure shear.

The buckle folds in Watkinson's experiments are very cylindrical, because the extension direction is parallel to the hinge lines (Watkinson 1975). In contrast, upright folds within the Vestranden gneiss complex can be somewhat non-cylindrical (e.g. Figs. 2 and 8). One explanation is that initially cylindrical folds may have been modified in both shape and geometry by a later period of deformation. Alternatively, the strain magnitude may have varied along strike, and this variation produced the deviation from cylindricity (see e.g. fig. 7–105 in Ramsay 1967). We believe that at least some of the non-cylindricity in the Vestranden folds is related to heterogeneous strain in a bulk pure shear (Fig. 17c).

Robinson & Krill (1991) have observed a similar sequence of deformational events to the southwest in the Surnadal synclinorium (Fig. 1). Trøndelag ophiolites were first thrust onto continental rocks, then folded, and finally subjected to very intense wrench-type sinistral shear.

Møre-Trøndelag fault system

The Verran valley is a first-order structural feature, separating distinctly different tectonic units, geometries and kinematics in the Vestranden gneiss complex to the north from the Inner Fosen cover–basement complex. Given the disparity in geometry and kinematics on either side, we expect a complex deformation history within this broad zone of accommodation. However, this transition seems to be dominated by a large, WSW-directed wrench component. We believe the penetrative deformation within this wrench zone is Scandian in age, but the relative timing with other first-order structures to the north and south is not clear. This sinistral wrench probably post-dates the early subhorizontal shearing and nappe stage, but is perhaps coeval with subhorizontal shortening. The magnitude of Scandian transcurrent movement along the Verran valley is not known; later brittle deformations must be restored first.

The branching relationship at depth with the Einarsdalen décollement is speculative (Fig. 13). Both deformation zones appear to be long-lived features, and probably did not develop sequentially. During main phase Scandian collision, the Einarsdalen décollement might have extended beneath the Verran valley wrench zone into Inner Fosen. Continuing the analogy with the southern Appalachians, this wrench zone might be comparable (in some aspects) to the Brevard zone (e.g. Bryant & Reed 1969). An early listric or stair-step thrust, placing rocks of Inner Fosen on the Vestranden gneiss complex, would then be reactivated as a strike-slip stretching fault separating different structural domains.

The Gullbergvatn, Hitra-Snåsa and Verran faults all show evidence for sinistral strike-slip on orogen-parallel shear zones, much of which took place under low-grade metamorphic conditions (see also Grønlie *et al.* 1991). These displacements are geometrically and kinematically similar to the wrench component scattered throughout the Vestranden gneiss complex, that formed at higher grade (typically biotite zone). Retrograde strike slip on the Møre-Trøndelag fault system could represent either a later phase of movement than the wrench in the Vestranden gneisses, perhaps further along the *P–T–t* path, or could be coeval with the wrench component, albeit at a shallower crustal level. Séranne (1992) has suggested that strike slip may post-date even late orogenic structures, such as the shallow W-dipping detachments that floor the Devonian basins (Hossack 1984, Séranne & Séguret 1987). We believe that sinistral strike-slip motion on steeply dipping structures initiated during the main phase of the Scandian collision and continued through late orogenic uplift, recognizing the need for more direct constraints on the slip history.

Inner Fosen basement–cover complex

Scandian structures on Inner Fosen are dominated by a basement-cored, W-directed, asymmetric fold nappe (Fig. 13). The fold axes for this large nappe are oblique to and fanned about the predominantly E-plunging mineral lineation (see Hull & Adamy 1989). Unlike folds in the Vestranden gneiss complex, we interpret this map-scale non-cylindrical fold as developing during progressive shear by hinge line rotation. This fold and associated kinematic indicators provide more evidence for generally W-directed transport in this portion of the Scandian hinterland. The direction of transport is difficult to define exactly, as the minor structures vary in geometry in the Leksvik map sheet.

We note a consistent clockwise rotation of mineral stretching lineations from the Vestranden gneiss complex (050°) through the Verran valley (060°) into Inner Fosen (074°). Transport directions in the Leksvik area are thus more orogen-oblique, in contrast to the orogen-parallel geometries in the Vestranden gneisses to the north, and the orogen-transverse lineations much further to the south and east (Hossack & Cooper 1986). Inner Fosen Peninsula thus represents a kinematic transition zone between the Vestranden hinterland and a more typical fold-and-thrust belt style in the cover-dominated sequences (e.g. Gilotti & Kumpulainen 1986).

Hinge lines of minor folds of schistosity are parallel to mineral lineations in the Leksvik area, akin to structures within the Vestranden gneiss complex, however the hinge planes show a variety of orientations dominated by shallow dips (Fig. 12). Shallowly inclined to recumbent minor folds of cleavage are common in the cover sequences of the central Scandinavian Caledonides (e.g. Roberts & Wolff 1981) and have been ascribed to late orogenic gravitational collapse. The polyclinal buckle folds in the Leksvik sheet are more consistent with subhorizontal constrictional strains, rather than vertical flattening; other minor structures indicative of vertical flattening, such as chocolate tablet boudinage, are lacking. A N–S pure shear superimposed on a large, W-directed simple shear will produce subhorizontal constrictional strains with E–W extension, suggesting that later subhorizontal shortening is also present in the Leksvik area, albeit with a different geometry than in the Vestranden gneisses.

COLLISION MODEL

The main characteristics of our study area are the early NW-directed thrusting and shear, related to crustal imbrication and high-grade metamorphism, followed by both orogen-parallel extension and orogen-parallel shear. We first briefly review some global models that are relevant to various aspects of this structural chronology, then conclude with a plate-scale model for the Scandian collision on our traverse.

One of the most popular models for orogen-parallel transport is slip partitioning (Fig. 18). Oblique plate convergence is generally not reflected in 'oblique' focal mechanism solutions for earthquakes at intra-oceanic trenches (e.g. Jarrard 1986). Instead, subduction zone earthquakes have P and T axes in a more down-dip orientation, normal to the strike of the trench. The 'missing' seismogenic component is taken up by strike-slip on steep faults parallel to the trench, often in the forearc region. This model of slip partitioning during oblique collision has subsequently been applied to various orogenic belts such as the North American Cordillera (Beck 1983, Avé Lallemant & Oldow 1988), the Alps (Mattauer 1986, Lacassin 1989) and the Caledonides of East Greenland (Holdsworth & Strachan 1991).

The mechanisms for partitioning of slip are not clear. Older faults, such as basin-forming normal faults, may be reactivated as strike-slip zones, or early thrusts may be rotated or folded into unfavorable orientations for further dip-slip movement. Body forces in the orogenic wedge may drive dip slip along thrusts, with the oblique convergence then 'balanced' by strike-slip motion. Woodcock (1986) asserted that oblique displacements in ductile crust are not partitioned into discrete strike-slip and dip-slip structures. We believe that the slip partitioning model is more applicable to deformation under low-grade, seismogenic conditions, perhaps late in the orogenic cycle.

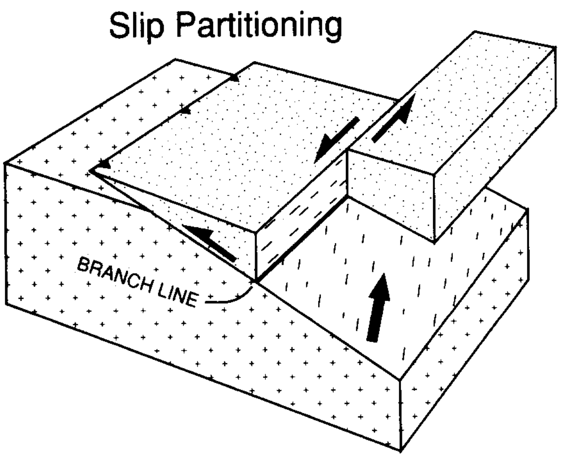

Fig. 18. Oblique transport can be partitioned into orogen-transverse and orogen-parallel components of displacement on thrust and strike-slip faults. The two sets of faults share a common detachment.

Orogen-parallel strike-slip faults need not require oblique convergence, however. An orogen-parallel transform can be generated by annihilation of a ridge crest at a trench (e.g. Atwater 1970, Beck 1986), even if convergence is orthogonal. Other types of strike-slip faults at plate boundaries are discussed by Woodcock (1986). Irregularities in the plate margins such as salients, recesses, and microplates of various sizes and shapes can lead to orogen-parallel strike-slip (e.g. Dewey *et al.* 1986, Mattauer 1986).

The topology or boundary conditions of plate interaction can play an important role in generating orogen-parallel transport during collision. The free plate boundary of the South China Sea allowed one-sided escape during essentially orthogonal continent–continent collision between India and Asia (e.g. Mercier *et al.* 1987, Peltzer & Tapponnier 1988). Much of the near-surface, N–S shortening and E–W extension in Tibet is localized along strike-slip faults such as the Altyn Tagh and Kun Lun faults. Orogen-parallel extension is also associated with N–S-striking grabens. The one-sided escape of Asia requires a dextral stretching fault along the southern border of Tibet, which has recently been documented by Pêcher *et al.* (1981). We think there are worthwhile analogies between the deeper levels of the India–Asia collision and the hinterland of the Scandinavian Caledonides, but the different scales preclude any direct comparison.

One possible configuration of plates, arcs and ocean basins leading to the Scandian orogeny is shown in Fig. 19. This configuration is both highly simplified, one of several permissible topologies, and relevant only to our particular study area. Iapetan rifting produced a number of continental microplates separated by oceanic crust, as opposed to a single ocean basin between Baltica and Laurentia. Each of these continental massifs was fringed by Late Proterozoic to Early Paleozoic rift-drift sequences and overlain by shelf strata of similar age. This model differs radically from most previous interpretations in that the basement rocks of the Western Gneiss region, Inner Fosen and Vestranden are not structurally continuous with the Baltic shield prior to continental collision (cf. Roberts & Wolff 1981, Gee & Sturt 1985).

There is good evidence for Late Cambrian to Early Ordovician imbrication of the Balto-Scandian continental margin (Fig. 19), with emplacement of both miogeoclinal and eugeoclinal sequences resulting in regional metamorphism, locally to very high grade (e.g. Gee & Sturt 1985). Partial uplift and cooling of this region below approximately 500°C took place in the Early to Middle Ordovician (460–470 Ma), based on ^{40}Ar–^{39}Ar data (e.g. Dallmeyer & Stephens 1991). This eastern portion of the orogenic belt was rejuvenated during the main Scandian event around 415 Ma (Claesson 1982).

Closing of the Iapetan Ocean basins and concomitant development of both continental margin and island arcs began in the Early Ordovician, and continued locally up

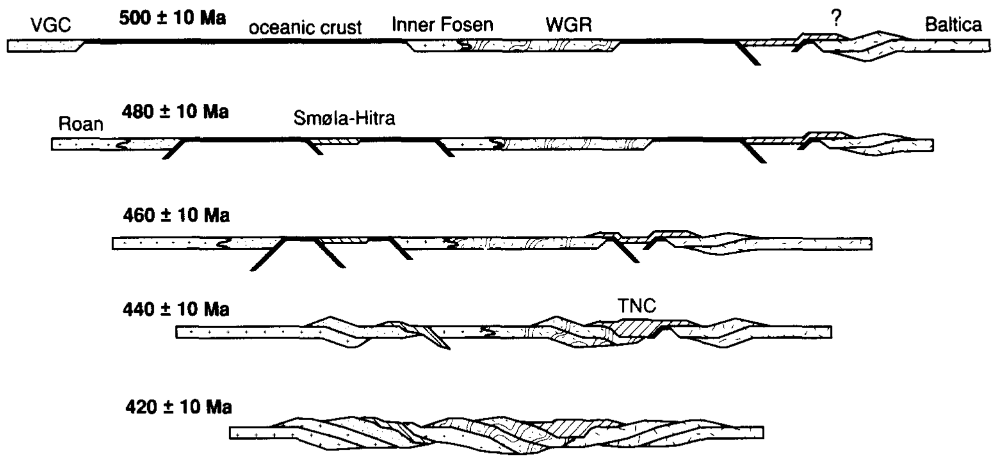

Fig. 19. NW–SE cross-section showing one model for plate interactions, illustrating southeastward underthrusting of the transect elements. Continental crustal segments, arcs (Smøla-Hitra and Trondheim nappe complex), and intervening ocean basins are shown.

to the Silurian (Fig. 19). The Trondheim nappe complex is a complicated amalgamation of Early Paleozoic arcs and ophiolites, locally containing Laurentian faunas (see Stevens & Gee 1985, Stevens 1988 for reviews). Subduction beneath Inner Fosen, Vestranden and Roan at various stages throughout the Ordovician is indicated by widespread, calc-alkaline plutonism (Möller 1988, Tucker 1988, Thorsnes *et al.* in review). The rather long-lived arcs of Smøla and Hitra contain Lower to Middle Ordovician fossils of Laurentian affinity and intrusions ranging in age from 460 to 430 Ma (Gautneb 1988, Tucker 1988). In Fig. 19, we show the Smøla-Hitra arc sutured between Vestranden and Roan gneisses to the northwest, and the Western Gneiss Region and Inner Fosen to the southeast, however this specific topology is not crucial to our model.

Southeastward drift of the continental fragments and arcs relative to the Trondheim nappe complex resulted in a series of continent–arc and continent–continent collisions (Fig. 19). The Trondheim nappe complex was thrust both southeastward onto the shortened Baltic margin and northwestward onto the Western Gneiss Region and other outboard microcontinents, resulting in a crustal scale pop-up (Gilotti & Hull 1991). Crustal doubling in the Western Gneiss Region led to regional high-grade metamorphism in the Silurian, however our imbrication geometry is different from most previous models (e.g. Andersen *et al.* 1991).

The northwestward transport of the Vestranden gneiss complex above the Einarsdalen décollement is related to similar southeastward underthrusting. Crustal imbrication produced regional, high-grade metamorphism, dated in both the Roan and Vestranden gneisses at 435 Ma (Schouenborg *et al.* 1991, Dallmeyer *et al.* 1992). Continued collision resulted in lateral extrusion of the Vestranden thrust sheet above the relatively rigid Roan block, accommodated by decoupling along the Einarsdalen décollement.

It is difficult to evaluate the variables that produced horizontal, orogen-parallel extension in the Scandian hinterland, rather than the vertical extension typically seen in the shallower, more foreland portions of fold and thrust belts. Extension within thickened crust has traditionally been attributed to an increase in the vertical compressive stress component due to build-up of topography. However, such a static mechanism does not account for the dynamic stresses in deforming bodies (England & Houseman 1989), and does not explain the simultaneous shortening and extension observed in the Vestranden gneiss complex. Subhorizontal extension would be favored by a decrease in flow strength (England & Houseman 1989), produced by thermal relaxation of tectonically thickened crust, or perhaps controlled by lithology and mineralogy.

In order to refine our structural and tectonic models, we must know more about the extent of orogen-parallel stretching and shear within the Caledonian hinterland, and have a better grasp on both the relative and absolute timing of deformation increments during Scandian collision. Syn- to late collisional along-strike slip has been recently noted from the northern part of the Western Gneiss Region (Robinson & Krill 1991) and from the northeastern portion of the East Greenland Caledonides (Holdsworth & Strachan 1991), however both the distribution and amount of orogen-parallel shear on either side of the collision zone is unknown. The quantitative evaluation of strain increments is necessary.

The magnitude of orogen-parallel extension in the Vestranden gneiss complex appears to be quite large and perhaps dominates the finite deformation. The structural style, with stretching parallel to hinge lines of major upright folds formed during regional contraction, is very distinct, but has been described only briefly from a few other orogenic belts (see e.g. Dewey *et al.* 1986, Ellis & Watkinson 1987, Hatcher & Tabor 1989). Extension parallel to orogenic strike may be an important deformation element in the hinterland of many other collisional orogens, and deserves much further investigation.

CONCLUSIONS

Heterogeneous shortening within the hinterland of the central Scandinavian Caledonides has been partitioned into shear zones bounding variably deformed structural domains. Three tectonic units on Fosen Peninsula in central Norway represent different structural levels, from high-pressure granulites in the north (Roan), through polyphase gneisses (Vestranden gneiss complex), to cover-dominated sequences in the south (Inner Fosen).

An orogen-parallel family of structures, produced by longitudinal extension and shear accompanying subhorizontal shortening, dominates the middle structural level. Shortening above a NW-directed décollement accommodated orogen-parallel extension in the thrust sheet; this style of deformation may be common in the deeper parts of orogenic hinterlands.

The shallowest crustal level on the transect shows a 'Pennine-style' of deformation, with a non-cylindrical fold nappe and associated minor structures indicating westward transport, somewhat oblique to the orogenic grain. The structures on Inner Fosen represent a transition to the orogen-normal transport typical of the shallowest crustal levels of the collisional foreland.

Transport towards the northwest and west on our traverse is in stark contrast to previous models for this region, which show either para-authochthonous crust or southeasterly translations. This 'backthrusting' represents a fundamental change in transport symmetry, related to underthrusting of Vestranden beneath the arcs and ophiolites of Trøndelag.

The crust is therefore stratified both geometrically and kinematically, however the larger scale tectonic setting that produced this heterogeneous deformation is still uncertain.

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