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# Strain softening induced ductile flow in the Särv thrust sheet, Scandinavian Caledonides

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**Abstract**—The external zone of the Scandinavian Caledonides is characterized by thrust sheets composed of sedimentary rocks derived from the continent Baltica. The Särv thrust sheet is unique amongst them because the 4.5–6 km thick, flat-lying Tossåsfjället Group clastics were intruded by a sub-vertical tholeiitic dike swarm prior to thrusting. The Särv thrust sheet is a large horse (100 km long and 80 km wide) with a classical ramp-flat geometry. The original intrusive relationships of the dikes are preserved throughout most of the thrust sheet and deformation is concentrated at the base in a mylonite zone several tens of meters thick. In this zone, meter-thick dikes rotate into parallelism with the basal thrust and concordant mylonites of both lithologies are intercalated on a centimeter to millimeter scale. The overall rigid block geometry of the Särv thrust sheet, and the intense strain gradient recorded by the rotation of the dike swarm and the concurrent development of mylonites in the thrust zone, suggest that thrust emplacement took place by ductile yield and on-going pseudoplastic flow. Continual strain softening localized ductile flow in this zone by means of grain-size reduction, reaction-enhanced ductility and grain-boundary sliding — processes inferred from the microstructures.

## INTRODUCTION

MECHANICAL models of thrust emplacement have received much attention in recent years and generally fall into two categories: those which employ a Mohr–Coulomb criterion for brittle failure (Hubbert & Rubey 1959, Davis *et al.* 1983) and others which favor yield governed by power-law flow (Elliott 1976, Chapple 1978). Brittle-failure models are formulated to require frictional sliding of a rigid block over its footwall. Ductile-yield models require some amount of penetrative deformation to have occurred within the thrust sheet and to increase in intensity towards the fault. Card deck models (Elliott 1976) qualitatively illustrate the strain distribution expected in an internally deforming thrust sheet and suggest that a large component of simple shear strain may operate at the base.

In most natural systems some penetrative deformation is observed, either distributed throughout the thrust sheet, commonly in the form of cleavage development (Mitra & Elliott 1980), or concentrated at the base (e.g. Schmid 1975, Mitra 1984, Wojtal this volume). We will describe an extreme case of strain localization in which flow, induced by strain softening, was concentrated in a mylonite zone which is less than 0.1% of the thickness of the thrust sheet. Two excellent examples of this process have already been described: the Lochseiten calc-mylonite at the base of the Glarus thrust sheet in the Swiss Helvetic Zone (Schmid 1975) and crystalline basement mylonites above the Linville Falls fault in the Blue Ridge Province of the southern Appalachians (Boyer 1978).

The Särv thrust sheet offers an opportunity to study concurrent mylonite development and thrust emplacement at upper greenschist grade conditions. As Schmid (1975) pointed out, care must be taken to demonstrate that mylonite formation is synchronous with and directly related to thrusting. In this light, our aim is to describe the geometry and microstructures of the basal deformation zone and relate them to the overall configuration and displacement history of the Särv thrust sheet. Details of the deformation mechanisms which produced the strain softening are under investigation and will be treated separately (Gilotti in prep.).

## GEOMETRY OF THE SÄRV THRUST SHEET

### *Regional setting*

The Särv thrust sheet is located in the southwestern corner of Jämtland County, Sweden between Östersund, Sweden and Trondheim, Norway (Fig. 1). This lies in the geotraverse area described by Gee (1975, 1978). The Särv thrust sheet consists of Upper Proterozoic clastic sediments derived from the Baltoscandian miogeocline and a tholeiitic dolerite dike swarm which was intruded prior to Caledonian deformation. This distinctive sequence of rocks was first defined as a separate lithotectonic unit (i.e. the Särv nappe) by Strömberg (1955, 1961); and it has subsequently been recognized throughout the central Scandinavian

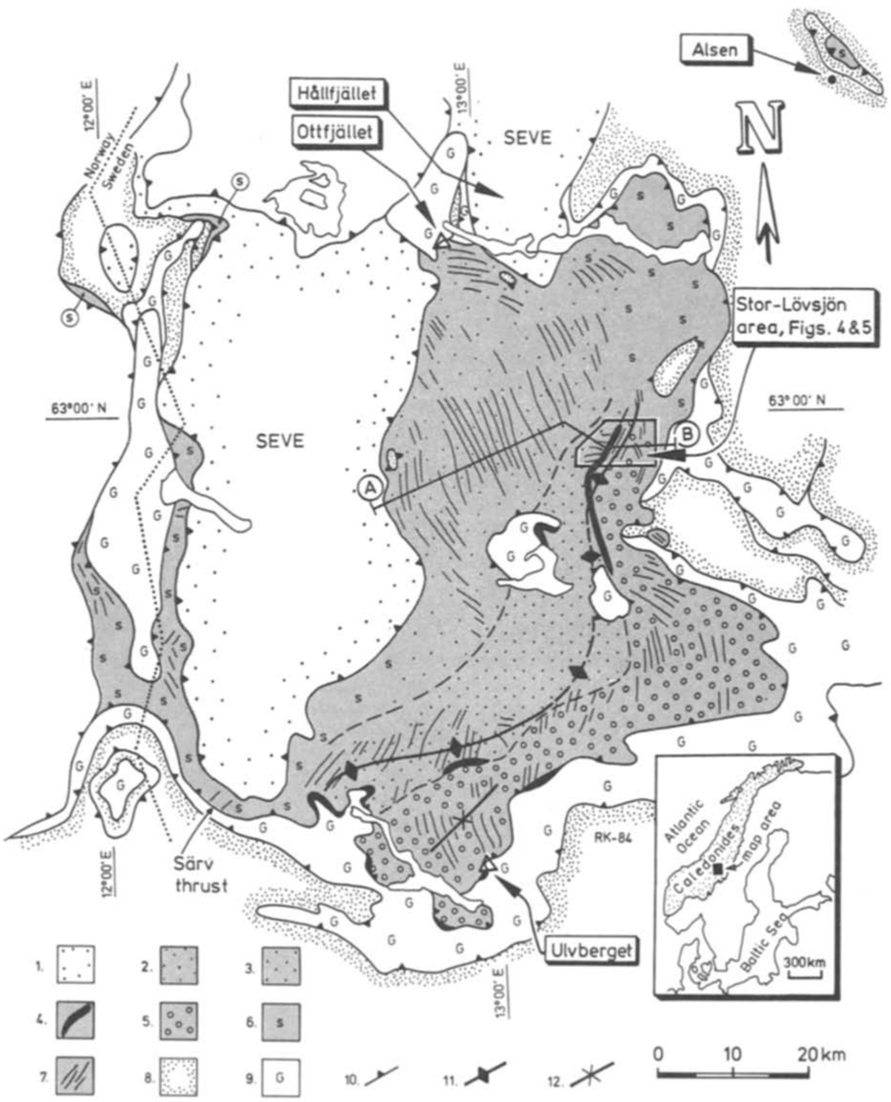


Fig. 1. Generalized geologic map of the Särv thrust sheet in the southern Scandinavian Caledonides (inset) showing the major lithotectonic units and thrusts. Symbols are 1, Seve Nappe; 2–7, Särv lithologies (2, Lunndörrsfjällen Formation; 3, Kråkhammaren Formation; 4, Storån and Lillfjället Formations; 5, Lövan Formation; 6, formation unknown; 7, dolerite dikes); 8, Baltoscandian cover rocks; 9, Baltoscandian crystalline rocks undifferentiated; 10, thrust fault; 11, anticline; 12, syncline. A–B, location of section line (Fig. 2).

Caledonides as far west as the coast of Norway and northwards into Norbotten (Gee & Zachrisson 1979, Andreasson & Johansson 1983).

In Scandinavia, lithotectonic units separated by thrust faults are traditionally referred to as nappes and the nappe volumes are named and correlated on the basis of stratigraphic and paleogeographic similarities. An alternative aproach is to consider the geometry of branching thrust fault networks; the thrust fault surfaces are given names and the overlying rock volumes or thrust sheets are named after the fault. These two approaches are not incompatible (Hossack in press), but the geometrical approach provides more information on the sequential evolution of the thrust belt. We will refer to the main exposure of Särv rocks in Jämtland as the *Särv thrust sheet* in order to emphasize its overall thrust geometry, and to distinguish this sheet from other occurrences of the same group of rock types (but probably not the same thrust sheet) which can be collectively referred to as the *Särv nappe*.

The Särv thrust sheet extends approximately 100 km N–S and 80 km E–W, an area comparable in size to the Moine thrust belt in NW Scotland or the Pine Mountain block in the southern Appalachians (Fig. 1). Highly deformed and metamorphosed Precambrian crystalline and cover rocks (see Claesson 1981) of the Seve nappe immediately overlie the Särv thrust sheet and are thought to have been carried piggyback by subsequent motion on the Särv thrust. To the west along the Norwegian–Swedish border the Seve rocks locally pinch out and a sequence of volcano-sedimentary units of the Trondheim nappe complex overlies the Särv thrust sheet (Gee *et al.* in press). The Seve and higher nappes comprised an overburden of 15–20 km above the Särv thrust: 15 km is estimated from a synorogenic surface constructed for the Trondheim to Östersund cross-section and extrapolated from conodont alteration values at the thrust front (Hossack in press); 20 km is inferred from upper greenschist grade assemblages (400°C max.) in the thrust zone, assuming a geothermal gradient of

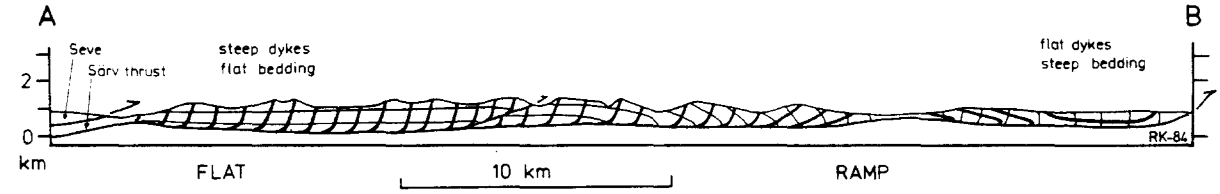


Fig. 2. Schematic profile AB (located on Fig. 1) through the Särv thrust sheet showing the orientation of dikes (as bold lines) and sedimentary bedding surfaces. The profile is drawn through the best exposures and is not in the transport direction.

20°C km$^{-1}$. The Särv thrust sheet was emplaced over another sheet of Baltoscandian derived cover sediments (the Offerdal nappe) and in places rests on Precambrian crystalline rocks (Claesson 1980) of the Tännäs augen gneiss nappe.

The stratigraphy of the Särv thrust sheet was established and the tectonic history discussed by Kumpulainen (1980). A brief summary of those results is included here.

In Late Proterozoic time, continental breakup and basin development led to the deposition of the Tossåsfjället Group, a thick clastic package dominated by feldspathic sandstone of which 4.5–6 km survives. Five formations are recognized (Fig. 1). The lowest of these is the Lunndörrsfjällen Formation, which consists of approximately 2000 m of feldspathic sandstone. This is overlain by another 2000 m of feldspathic sandstone and minor shale of the Kråkhammaren Formation. These rocks grade up into the 100 m thick Storån Formation, which is composed of dolomite and arenaceous dolomites. Tillite of the Lillfjället Formation unconformably overlies the dolomite and is at least 120 m thick. The contact between the tillite and the underlying dolomite is sharp; the contact with the overlying 1500 m thick Lövan Formation is gradational. Approximately 100 m of shale rapidly coarsens upwards to feldspathic sandstone and conglomerate typical of a major portion of the Lövan Formation. The top of the Lövan Formation is never visible, nor is the contact between the Lunndörrsfjällen and Kråkhammaren Formations. Based on lithostratigraphical similarity of the Storån–Lillfjället Formations to other carbonate–tillite associations in Scandinavia and Rb–Sr age determinations of 653 ± 7 Ma (Sturt *et al.* 1975) for tillites in Finnmark, the Lillfjället Fm. is correlated with deposits of the Varanger Ice Age in their type area in northernmost Norway (Kumpulainen 1981, Kumpulainen & Nystuen in press).

As rifting continued, the sedimentary sequence was intruded by a tholeiitic dike swarm, the Ottfjäll dolerites. A Rb–Sr whole rock isochron places the intrusion age of the dolerites at 720 ± 260 Ma (Claesson 1976). A recent $^{40}Ar/^{39}Ar$ date of 665 ± 10 Ma (Claesson & Roddick 1983) is favored because it is in better accordance with the tillite age in northernmost Norway. The dikes intruded undeformed sediments at high (70–90°) angles in a swarm that was approximately parallel to the N–S regional strike of the Baltoscandian continental margin (Kumpulainen 1980). Strömberg (1969) estimated that the Ottfjäll dolerites represent approximately 35% extension of the Tossåsfjället Group sediments. The original high angle between bedding in the sediments and the Ottfjäll dikes is preserved throughout most of the Särv sheet. No subsequent deformation occurred until the Caledonian Orogeny when the Särv sheet was thrust southeastwards onto Baltica. The areas where the Särv is thrust over the older Tännäs augen gneiss basement rocks need not represent out-of-sequence thrusting. More probably, the thrusts did not cut a layer cake stratigraphy; rather, the Tossåsfjället Group sediments were deposited in rift basins bounded by basement blocks (Kumpulainen & Nystuen in press). A zone of intense ductile deformation, of the order of 10–100 m thick, is found at the base of the Särv thrust sheet. The deformation zone represents less than 0.1% of the 4.5–6 km, pre-erosional thickness of the thrust sheet. In this basal zone, the dolerites are rotated into parallelism with bedding in the feldspathic sandstone and both rock types are mylonitized (Fig. 2).

Displacement estimates for the Särv thrust sheet range from a minimum of 80 km (i.e. the width of the sheet) to 100 km, determined from branch line maps (Hossack 1983), to greater than 250 km (Gee 1978, Kumpulainen & Nystuen in press) if all the components of the Särv nappe are restored as a single sheet to a pre-Caledonian basin position off the present coast of Norway. The 250 km figure overestimates the displacement on the basal thrust of the Särv sheet in Jämtland because the estimate includes the displacement accumulated by motion on later and lower thrusts which carried the Särv sheet piggybank onto Baltica.

*Internal sheet geometry*

Like many of the thrust sheets in the Scandinavian Caledonides, the Särv has a characteristic lens shape. Some authors (e.g. Gee 1978, Kautsky 1978, Andreasson & Johansson 1983) have interpreted such lenses as megaboudins formed during the gravitational collapse of an already established nappe pile. Others (e.g. Boyer & Elliott 1982, Hossack 1983, Lisle 1984) believe that lens shapes arise from branching fault networks. Lenses bound by fault surfaces are commonly formed in thrust belts when a younger flat thrust splays off the footwall of a higher and older thrust before eventually ramping upsection in all directions to rejoin the upper thrust surface. The line of intersection, termed a branch line (Boyer & Elliott 1982), forms a closed loop around the lens volume or horse. Hossack (1983) has used branch line maps to illustrate that the combined Tännas–Särv–

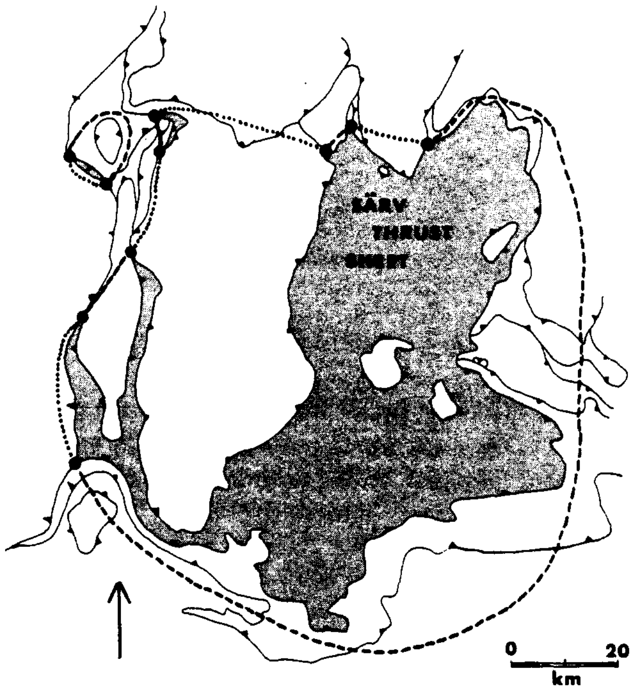


Fig. 3. Branch line map of the Särv thrust sheet; dashed lines represent the eroded portions and dotted segments represent buried portions. The branch lines intersect the ground surface at the large dots. The closed loop pattern of the branch line indicates that the Särv thrust sheet has the geometry of a horse.

Offerdal thrust system forms a huge horse in the central Scandinavian Caledonides. A similar analysis reveals that the Särv thrust sheet in SW Jämtland is actually a smaller horse within this larger structure.

The Särv thrust sheet thins from its present erosional thickness of 1500 m in the central regions of the sheet to zero at the Norwegian–Swedish border (Fig. 2). One can trace this zero isopach (i.e. branch line) in a N–S direction along the western edge of the map area by locating the intersections (i.e. branch points) of the Särv thrust with the overlying thrust (Fig. 3), which is in most cases the basal thrust to the Seve nappe. Some segments of the trailing branch line of the Särv sheet are eroded while others are clearly buried (Fig. 3). The trailing branch line bends sharply at the NW corner of the sheet where it passes laterally under the Seve nappe. The branch line continues in an ESE direction for about 15 km where it emerges and is eroded for the rest of its length. The Seve and the Särv thrusts join the overlying Köli thrust at leading branch lines in the foreland (Hossack 1983); and as no Särv rocks are found east or immediately south of those shown in Fig. 1, the Särv branch line must close on itself to form a loop, which is indicative of a horse.

An E–W profile through the sheet (Fig. 2) shows that the dikes and bedding in the sediments are approximately perpendicular throughout. In the central and western regions, the Tossåsfjället Group sediments are sub-horizontal and right way up and the Ottfjäll dolerites are sub-vertical, except within 200 m of the base of the Särv thrust sheet where the dikes curl to become concordant with the thrust. Eastwards and stratigraphically upsection, the sediments steepen and are locally overturned at Stor–Lövsjön (Figs. 1 & 4) in the hinge area of a large anticline. Imbricate thrusts, characteristic of other Baltoscandian cover sheets at this structural level

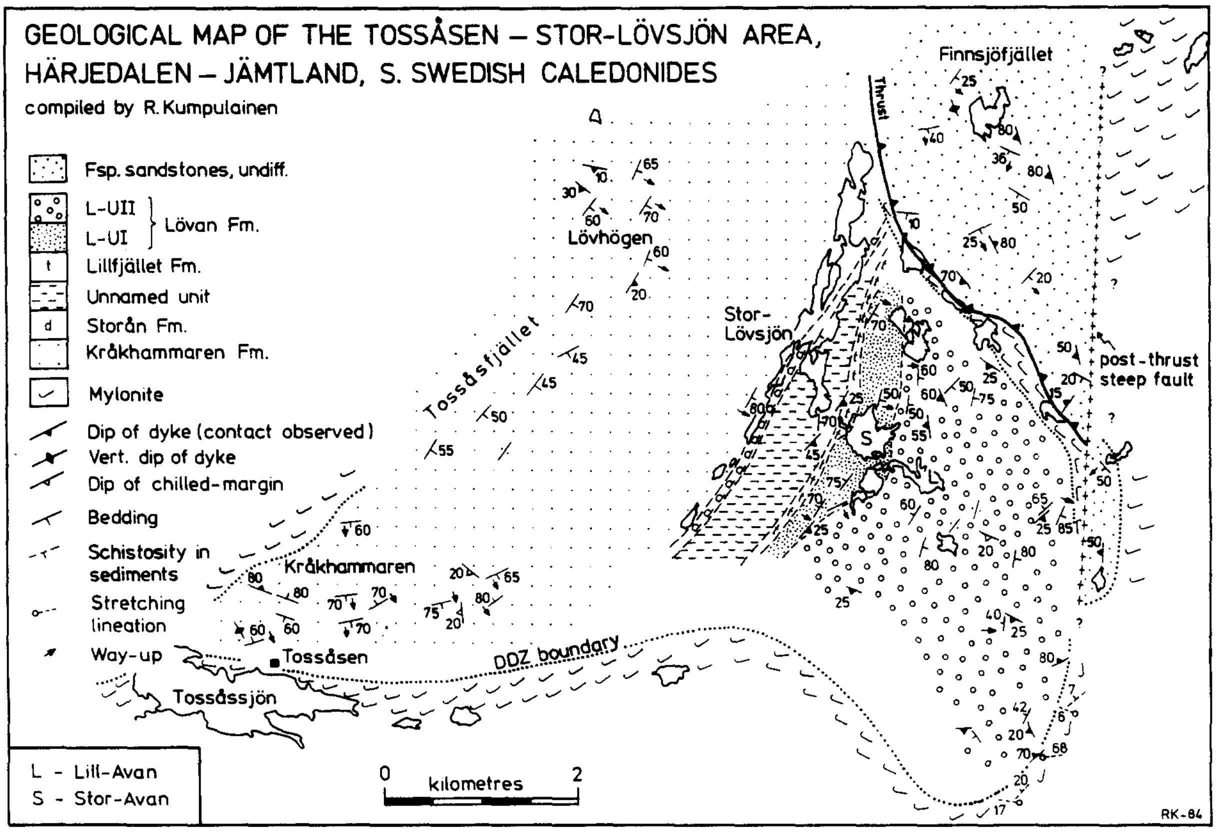


Fig. 4. Detailed geologic map of the Tossåsen–Lövsjön area in the hangingwall anticline at the leading edge of the Särv thrust sheet (location shown on Fig. 1). Permission for distribution approved by the Security Officer. The National Land Survey of Sweden 1985–10–04.

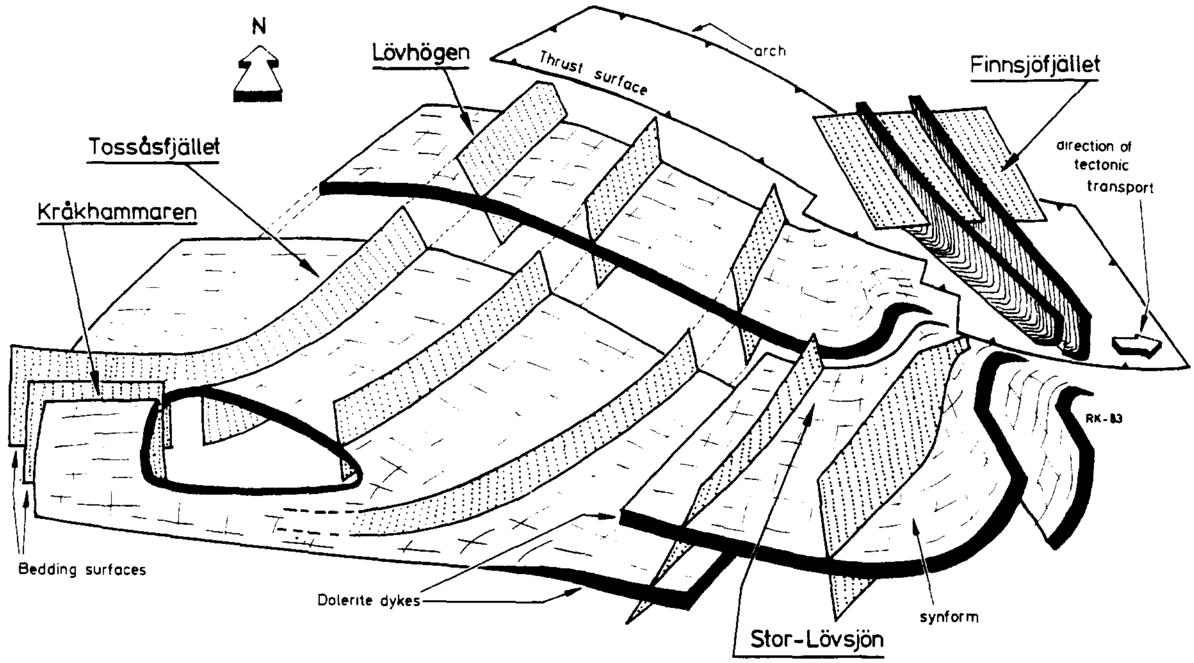


Fig. 5. Schematic diagram showing the structural relationships of the dikes and bedding surfaces in the hangingwall anticline and the geometry of the imbricate thrust in the Stor Lövsjön area. This diagram corresponds to the map data in Fig. 4.

in the Caledonides (Hossack in press), appear to be conspicuously minor features in the Särv thrust sheet. Where mapped, the imbricate thrusts exhibit relatively small displacements and little stratigraphic separation in comparison to the sole thrust.

The three-dimensional configuration of the dikes and sediments in the hinge of the anticline is shown schematically in Fig. 5 and corresponds to the detailed map of the Stor–Lövsjön area in Fig. 4. A number of important features are illustrated. (1) During folding, the dikes rotated in a passive manner so that the primary intrusive angular relationship with the sediments is preserved even in the hinge area. (2) Although bedding is locally overturned, a completely inverted recumbent limb is not present. Instead, bedding rotates abruptly into parallelism with the basal thrust (not shown on Fig. 5) in the same way that the dikes are rotated and truncated by the imbricate thrust at Finnsjöfjället. (3) The dikes in the hangingwall of the Finnsjöfjället imbricate slice have a brittle–ductile type shear zone geometry (Ramsay 1980). (4) The Finnsjöfjället imbricate slice was apparently emplaced first and then carried piggyback by the Särv thrust; subsequent motion on the Särv thrust has gently arched the pre-existing Finnsjöfjället thrust surface. The dikes in the main sheet are warped in a gentle synform, which is probably an accommodation structure formed during motion on the Särv thrust. (5) The root of the Finnsjöfjället thrust is unknown; however, the leading edge of this imbricate fault intersects either the Särv thrust where it ramps upwards, or the Seve–Köli thrust system at a branch line.

The profile (Fig. 2) shows that the basal thrust to the Särv sheet had a ramp-flat geometry. The thrust followed a long flat in the Lunndörrsfjällen Formation, ramped upsection at a low angle in the Kråkhammaren Formation, and then propagated as a steeper ramp through the Storån–Lillfjället–Lövan Formations. The hanging-wall anticline at Stor–Lövsjön formed when the thrust sheet reached a footwall flat. The trace of the axial surface of this ramp anticline approximately coincides with the current exposure of the carbonate–tillite contact (Fig. 1) along both the leading edge and the southern oblique ramp of the Särv thrust sheet.

Another ramp anticline occurs along the northern boundary of the sheet where the sediments are near vertical and the dikes trend E–W (Fig. 1) with gentle southward dips. Holmquist (1894) noted these 'peculiar' structural relationships at Ottfjället (Figs. 1 and 6). Strömberg (1961) mapped the traces of the dikes over the entire area of Fig. 1 and attributed the change in orientation to rotation of parts of the sheet. Paleomagnetic evidence (Bylund & Zellman 1980) also suggests counter-clockwise rotation of the dikes in the Ottfjäll area through 100–110° about a vertical axis during

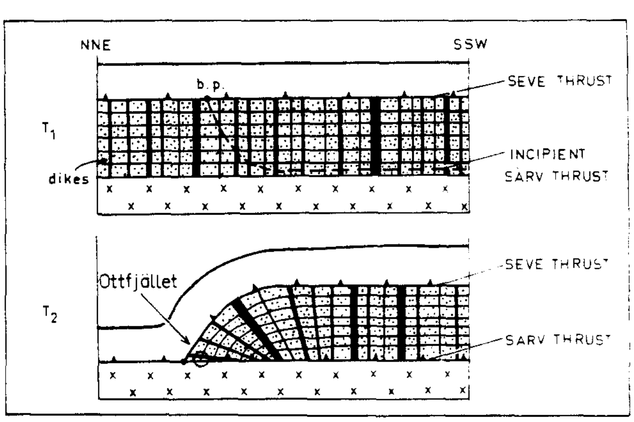


Fig. 6. Hangingwall sequence diagram showing the evolution of the lateral ramp at the northern edge of the Särv thrust sheet looking ESE in the transport direction. At Time 1 the Seve is in place and the incipient Särv thrust meets the Seve thrust at a branch point (b.p.). At Time 2 the Särv sheet has reached a flat where the ramp anticline forms.

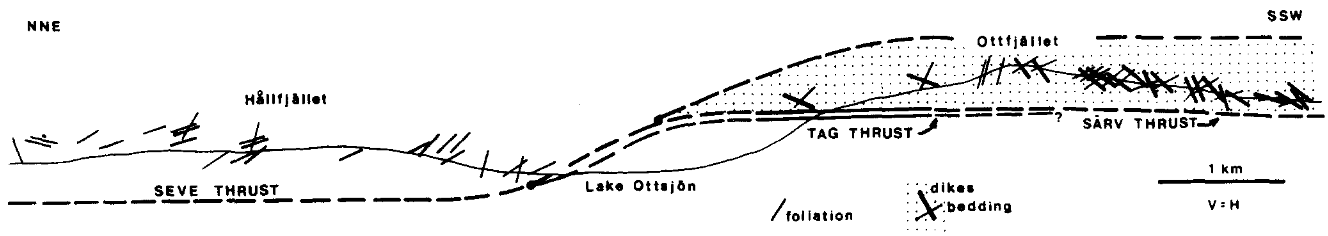


Fig. 7. Longitudinal section showing thrust sheet geometries in the lateral ramp at Ottfjället. Traces of the thrusts are approximately located with dashed lines. In the Särv sheet, bedding becomes steeper and dikes become shallower in the hinge area of the lateral-ramp anticline. Metamorphic foliations in the Seve schists and amphibolites steepen towards the SSW where the Seve sheet has been folded by motion on the underlying Särv thrust.

Caledonian deformation. The rotation probably represents confined flow along a cross-fault to a lateral ramp. Eventually, the Särv thrust cut up section laterally; when the sheet translated onto a flat the anticline formed automatically. The evolution of the lateral ramp at the northern edge of the sheet and its accompanying ramp anticline is shown schematically in a hangingwall sequence diagram (Fig. 6). A corresponding longitudinal section shows the projected traces of structural data (Fig. 7). The structure bears a striking resemblance to Harris's (1970) longitudinal section of the Pine Mountain block in the southern Appalachian foreland.

Because branch lines mark the intersection of a lower splay with an upper fault surface, they also represent the position and shape of footwall ramps (Hossack 1983). Symmetric branch lines connected to a leading branch line mark lateral ramps which by definition form in the transport direction (Elliott & Johnson 1980). Using this reasoning, Hossack (1983) obtained a slip vector of 155° for the combined Tännäs–Särv–Offerdal nappes. The ramp along the northern boundary of the Särv sheet suggests a transport direction of 100°. A more south-easterly slip vector would require a thrust cutting down-section across this ramp which is incompatible with the lateral ramp observed at Ottfjället. Unfortunately, the southern branch line of the Särv sheet is not well constrained; though the southwesterly trend of the anticline (trace shown on Fig. 1) suggests an oblique ramp (Hossack 1983) along this boundary of the horse. The footwall ramp to the Särv thrust is not the same structure defined by Hossack (1983) and implies that either the details of his analysis are incorrect due to the small scale (1:1,000,000) and preliminary nature of his base maps, or the displacement direction of thrusting changed from 100–120° to 150–160° with time. The latter is consistent with a tectonic model requiring oblique closure of the Iapetus Ocean and consequent underthrusting of Laurentia by Baltica (Hossack & Cooper in press).

*Basal Särv ductile deformation zone*

At the base of the Särv thrust sheet, there is an abrupt but continuous transition from relatively undeformed dolerites and sandstones to intensely deformed mylonites. The dikes record very high simple shear strains as they rotate as much as 90° over distances as small as 10 m into the mylonite zone above the thrust (Fig. 2). Although the geometry at the base of the Särv is identical to half of a classic shear zone (Ramsay & Graham 1970) and simple shear can account for most of the deformation, the possibility of some component of extension or compression perpendicular to the thrust cannot be excluded. Therefore, we follow Mitra (1978, 1984) and refer to the basal Särv thrust sheet as a ductile deformation zone (DDZ).

We will define the upper DDZ boundary as that surface which separates macroscopically unfoliated sandstones and dolerites from their strongly foliated equivalents in the DDZ beneath. In reality, the upper DDZ boundary has some thickness because the development of a foliation is dependent on the strain gradient and by its very nature is transitional. Above the upper DDZ boundary, the dike-bedding angles remain large, sedimentary structures are preserved and cleavage is rare. Within the basal Särv DDZ, very thin bands of altered dolerite are now parallel to the mylonitic foliation which is regionally concordant with the Särv thrust. All the primary sedimentary features are destroyed and the rocks are LS tectonites.

The upper DDZ boundary is such a sharp, striking feature that Strömberg (1961) placed the base of the Särv thrust sheet along it and included all the tectonites in a separate sheet which he called the Mylonite nappe. However, as there is no loss of structural continuity across the DDZ boundary, we place the Särv thrust in the 'mylonite nappe' at the point where obvious dolerite-derived schists disappear. Unfortunately, the base of the Särv thrust sheet lies mostly in lowland and bogs and the thrust contact is poorly exposed. Nevertheless, it should be possible for future workers to distinguish between Särv mylonites, and the K-feldspar bearing Tännäs augen gneiss mylonites and mylonitized Offerdal flagstones common in the footwall.

A complete section through the basal Särv DDZ is exposed at the northern border of the sheet on the west flank of Mt. Ottfjället (see Fig. 1). This section is located in the area of the lateral-ramp anticline where dikes are sub-horizontal and strike E–W; the sedimentary sequence is near vertical and youngs northwards (Figs. 6 and 7). At Ottfjället, typical Särv lithologies are thrust over approximately 40 m of K-feldspar augen bearing mylonites which look identical to mylonitized Tännäs augen gneiss (Röshoff 1978, Alm *et al.* 1980) found beneath the Särv thrust in the southern exposures of the sheet. A thin veneer of quartzite mylonite and biotite schist separates the Tännäs augen gneiss mylonites from

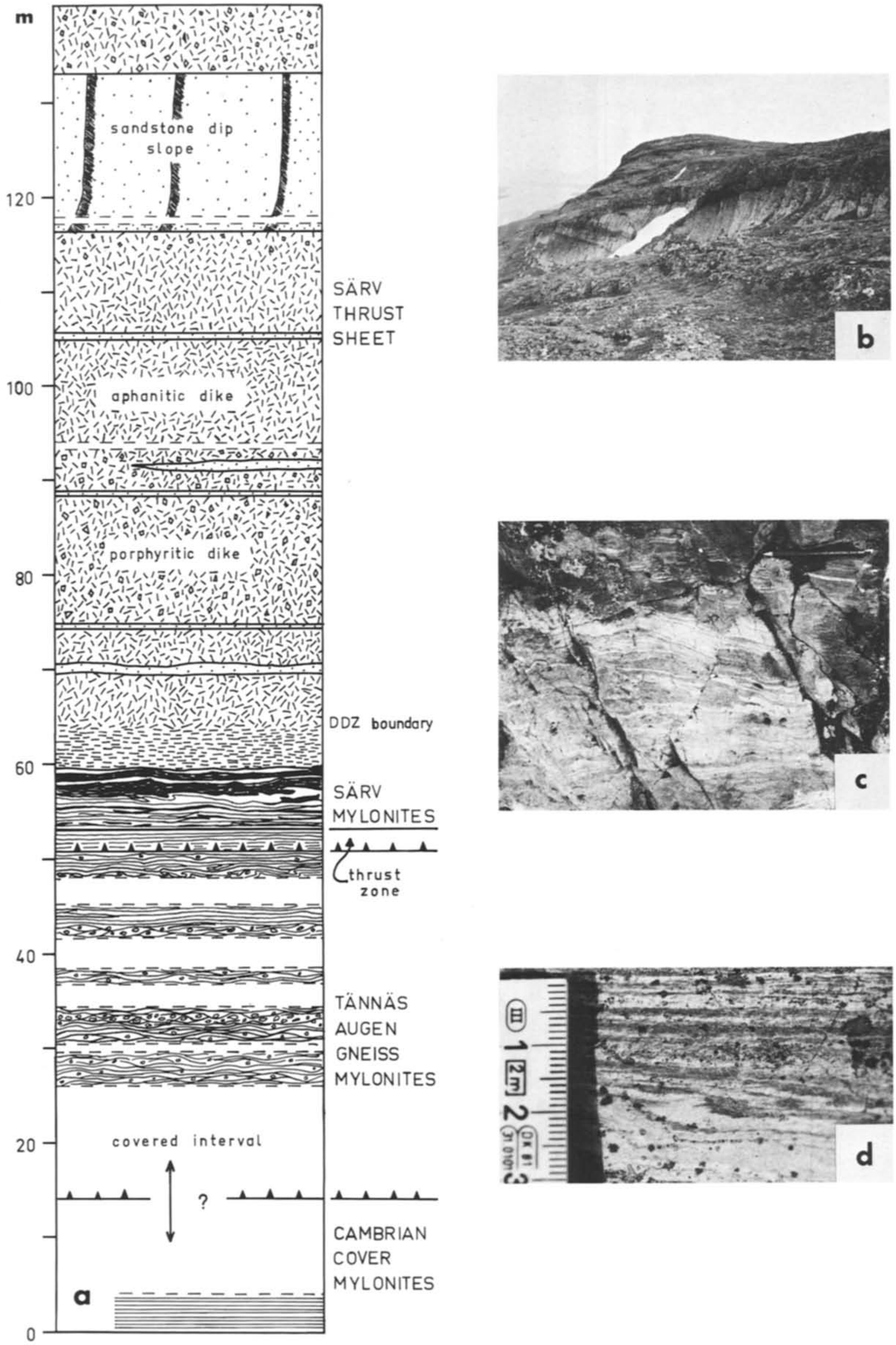


Fig. 8. (a) Measured section through the Särv DDZ at Ottfjället. (b) The structural relationships above the DDZ in the lateral ramp at Ottfjället; dikes are sub-horizontal and bedding surfaces (visible as dip slopes) are sub-vertical. (c) A pencil marks the contact of the lowermost dike in the measured section with the Särv mylonites beneath. The photo shows the banded nature of the mylonites and typical shear folds with flame structures. (d) Interbanding and folding of the metabasalt and quartzofeldspathic mylonites on a millimeter scale.

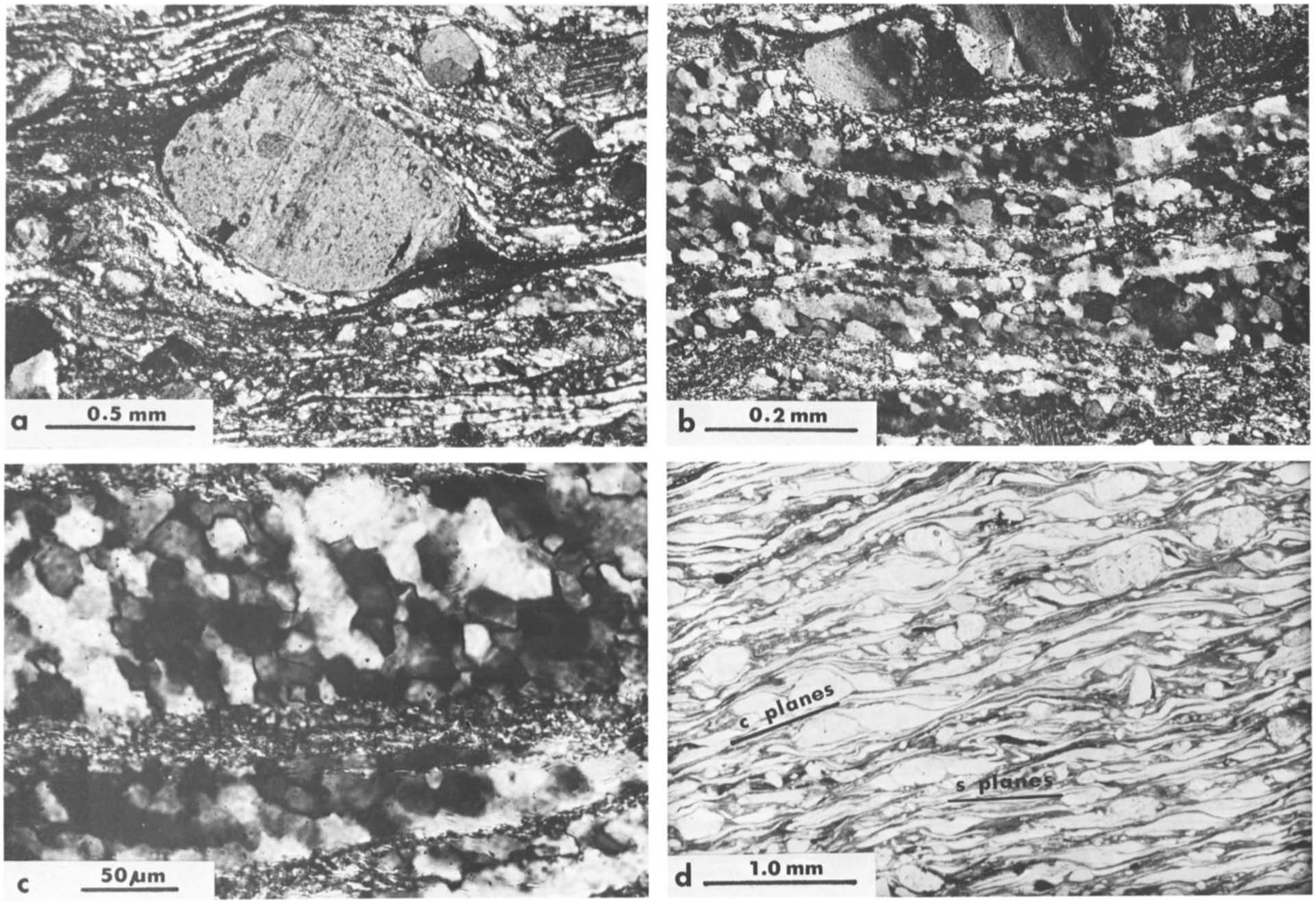


Fig. 9. Shear-sense indicators in the quartzofeldspathic mylonites viewed looking south record a sinistral sense of shear. (a) Asymmetric tails of recrystallized quartz (dark) in pressure shadows around *feldspar grains give* a *sinistral* sense of shear. (b) Long, straight sections of quartz ribbons are composed of dynamically recrystallized grains. (c) Shear sense is given by the oblique orientation of the long axes of the recrystallized quartz grains with respect to the ribbon boundaries (60 and 20° in photo). (d) Quartz ribbons form intrafolial folds in domains separated by parallel sets of fine-grained phyllosilicate-rich shear planes (C planes) which cut the mylonitic foliation (S planes) obliquely at 20°. Vergence of folds suggests overall sinistral shear sense.

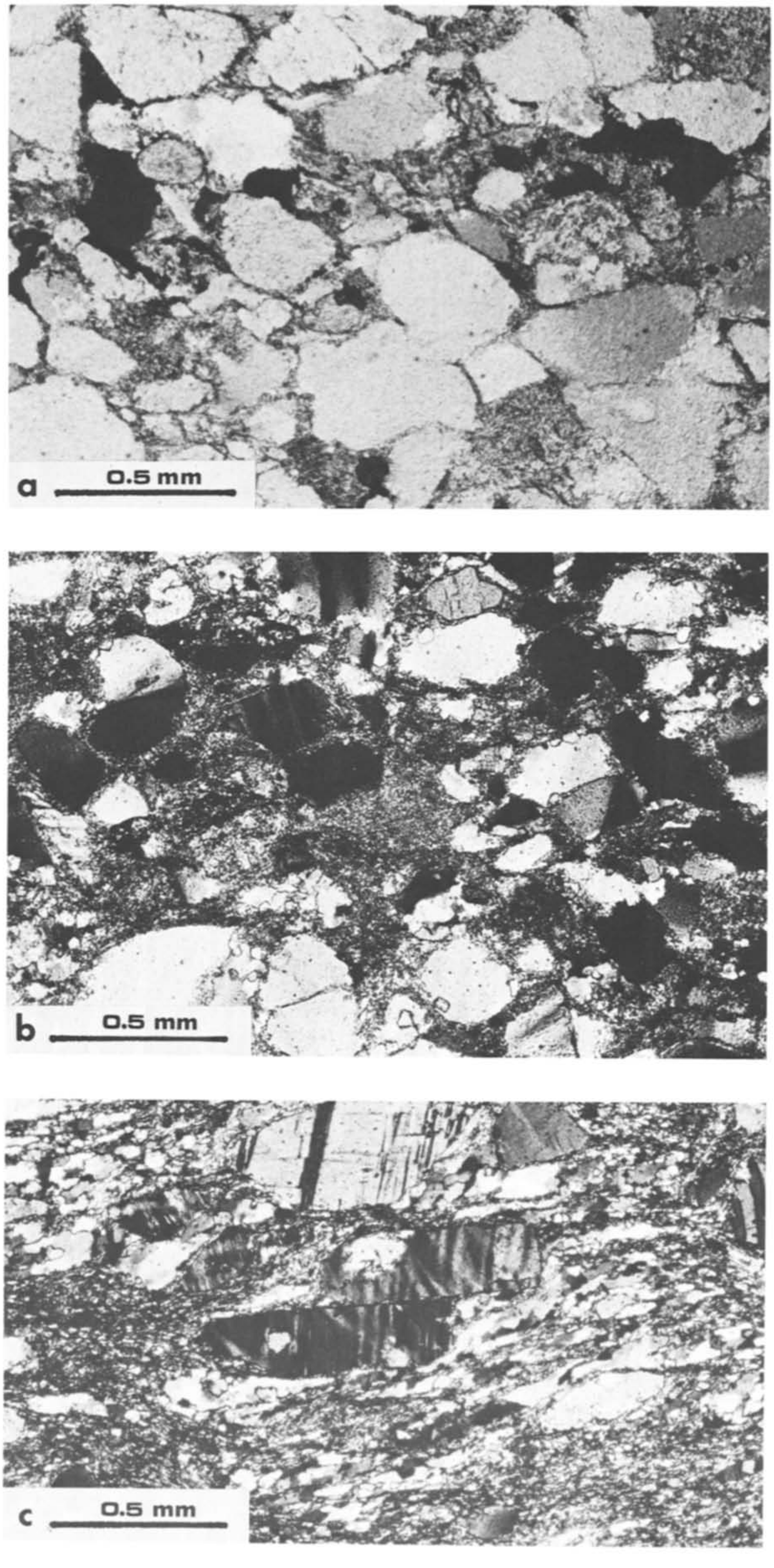


Fig. 11. (a) Sandstone protolith 10 m above the DDZB at Ulvberget (Fig. 10) shows sub-angular quartz and feldspar grains with low aspect ratios, low matrix percent, and some pressure solution (plane light). (b) Sandstone protolith viewed under crossed nicols shows significant plastic deformation. Many of the quartz grains have totally recrystallized but no foliation has developed. (c) Mylonite located 3 m beneath the DDZB. Dynamic recrystallization is complete; micas and quartz in the groundmass define a foliation. Grain boundary sliding has occurred along a fracture in the large, central feldspar.

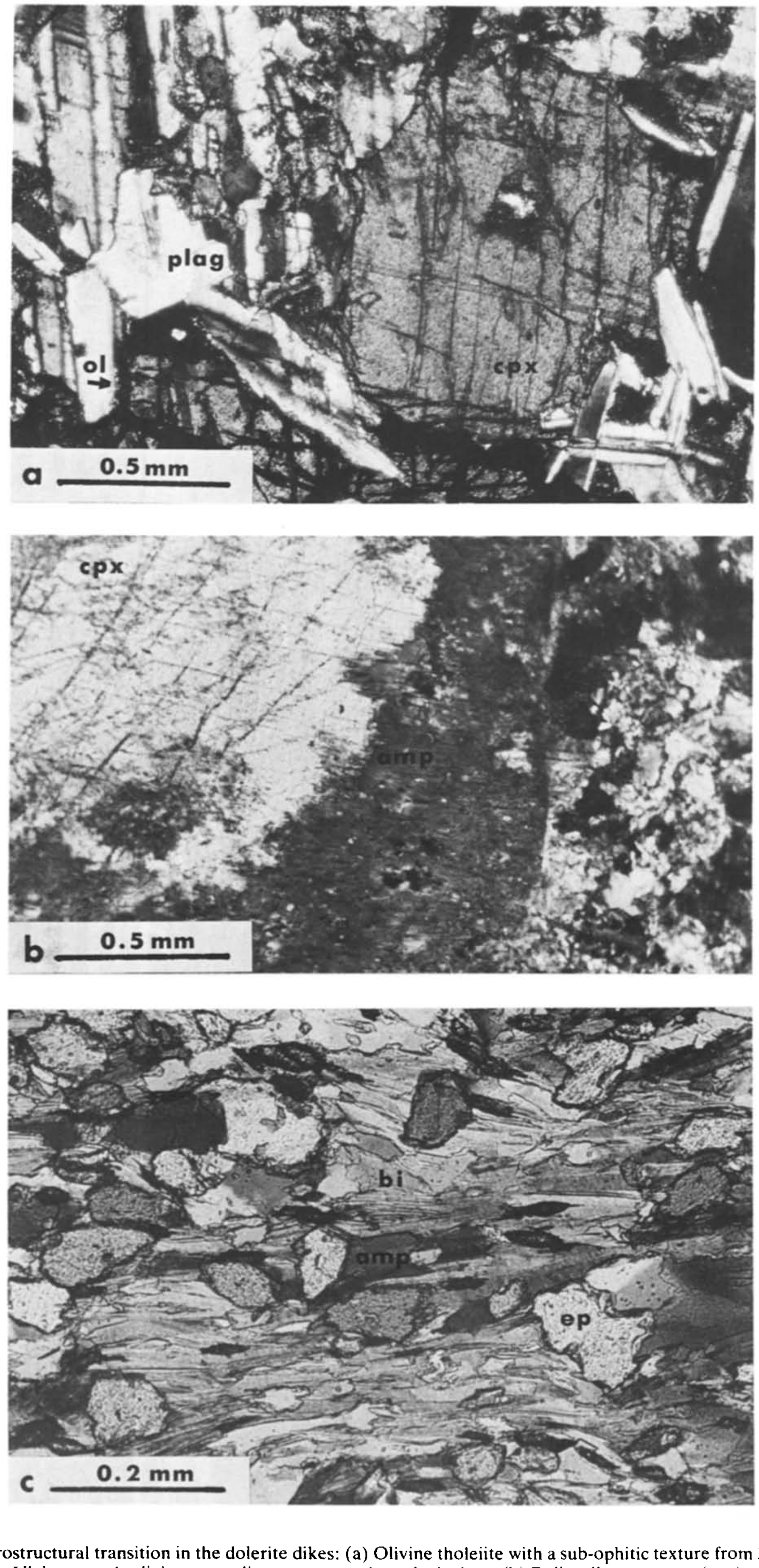


Fig. 12. Microstructural transition in the dolerite dikes: (a) Olivine tholeiite with a sub-ophitic texture from 30 m above the DDZB at Ulvberget. ol, olivine; cpx, clinopyroxene; plag, plagioclase. (b) Relict clinopyroxene (cpx) armoured by amphibole (amp) in a sample 3 m above the DDZB at Alsen (see Fig. 1 for location). Note the lack of a foliation in the groundmass. (c) Typical epidote – amphibolite schist in the DDZ at Alsen. This assemblage commonly occurs as bands only a few grains thick in the quartzofeldspathic mylonites. ep, epidote; bi, biotite; amp, amphibole.

rhyolite porphyry in the Mullfjället antiform to the west. The quartzites and schists are interpreted as highly sheared Cambrian cover rocks above the Baltoscandian crystalline basement.

The top of the measured section (Figs. 8a & b) shows the undeformed dikes and sediments 65 m above the Särv thrust: even here the dike-bedding angle is 84° and cross-stratification is ubiquitous in the sandstones. The 50 m of section above the DDZ boundary is dominated by aphanitic and porphyritic varieties of massive dikes separated by 0.5–2.0 m thick, subhorizontal lenses of unbedded quartzite. The DDZ boundary is located in the middle of the lowest undeformed dike (Fig. 8a) where a macroscopic foliation suddenly appears; 5 m below the DDZ boundary typical Särv mylonites are developed.

Although offset by late high-angle extension fractures, the contact between the lowermost dike and the mylonites is intact and continuous (Fig. 8c). In the mylonites, massive dikes are thinned and altered to schistose, cm-thick bands of upper greenschist facies assemblages. The mylonitic layering varies unsystematically from planar to anastomosing. Isoclinal and sheath folding have produced an interbanding of the greenschists and the quartzo-feldspathic mylonites on a cm to mm scale (Fig. 8d). In addition, many of the folds are restricted to individual layers (Fig. 8c), a characteristic of folding under the conditions of progressive simple shear. Boudins of dolerites are commonly found just above and within the DDZ (Fig. 10). As the mineralogy of the original dolerites alters to epidote–amphibolite schists, the competency contrast between the dikes and sediments diminishes and reverses itself in the mylonites where cusp and flame relationships indicate that the quartzo-feldspathic rocks behave as the more competent layers in the folds (Fig. 8c).

Preliminary observations suggest that the thrust contact between the Särv mylonites and the Tännäs augen-gneiss mylonites can be constrained to a ductile zone of finite width. The Särv and the augen-gneiss mylonites are separated on the basis of field criteria: both mylonite types can be strongly banded, but porphyritic greenschists are characteristic of the Särv mylonites while K-feldspar augen are taken as signatures of the basement gneisses. In the Ottfjället section, we have identified the Särv thrust as a zone 3 m thick (Fig. 8a) where the two mylonite types cannot be distinguished in the field and microstructural observations are currently lacking. Although further study of the textures and composition will certainly constrain the width of the fault zone, the important observation is that a brittle fracture separating the Särv thrust sheet from the Tännäs augen gneiss nappe is absent, and yet a minimum of 80 km of displacement is required in this thrust zone to juxtapose rocks with dikes over rocks without dikes. Eighty kilometers is simply the width of the Särv sheet measured in the displacement direction along the lateral (north) branch line.

We know from the ramp-flat geometry of the Särv thrust sheet that the basal thrust is a marked structural discontinuity. The mechanism of emplacement for the Särv sheet must have involved considerable ductile flow in this 10 m (at Ottfjället) to 100 m (inferred) thick mylonite zone for a large part of its movement history. Late-stage brittle features (e.g. cataclasites parallel to the mylonitic foliation) are not commonly found in this zone. This implies that the Särv thrust was not active at shallow crustal levels and that displacement was transferred to later, lower faults which carried the Särv sheet to higher crustal levels.

## MICROSTRUCTURES

### *Shear indicators*

Microstructures in the mylonites preserve a deformation fabric which indicates that simple shear accounts for a large component of the total strain in the DDZ. As previously noted, the mylonites are LS tectonites; the lineation is developed as a quartz-aggregate stretching direction lying in the schistosity plane defined by the phyllosilicates. Because the mylonites are so fine-grained, lineations are often difficult to measure in the field; however, Strömberg (1961, plate III 36) shows a stereoplot of stretching directions from 119 localities in the basal Särv mylonites and the augen gneisses. The stretching lineations range in orientation from E to S with a maximum trend toward the SE and are consistent with the general displacement direction of the Särv thrust inferred from the regional geology.

Shear-sense indicators (Simpson & Schmid 1983) are common in the Särv mylonites and also concur with the displacement direction of the Särv thrust towards the southeast. Asymmetric pressure shadows around porphyroclasts occur in both the greenschist and the quartz–feldspar–mica mylonites. In the greenschists, complicated growths of quartz, calcite and phyllosilicates occur in the shadows of saussuritized plagioclase phenocrysts and amphibole augen. Quartz tails occur around all sizes of feldspar clasts in the quartz–feldspar–mica mylonites (Fig. 9a). The asymmetry of the tails in both lithologies consistently records a top-to-the-southeast sense of shear.

In many cases, subgrain orientations in the quartz ribbons preserve an optically visible, dynamic recrystallization fabric which indicates the sense of shear (Law *et al.* 1984). Figure 9(b) shows a long straight section of several typical quartz ribbons; the boundary of the ribbons is marked by thin micaceous layers which define the foliation in the mylonites. The quartz ribbons have an internal fabric defined by grain elongation and the alignment of new grain boundaries which is at acute angles to the ribbon boundaries (Fig. 9c). Preliminary optical observations indicate that the oblique angle varies unsystematically between 0–60° from ribbon to ribbon in the DDZ. Quartz *c*-axes for the subgrains plot along an asymmetric girdle which is consistent with the sinistral sense of shear shown in the photograph (Fig. 9c). This suggests that different quartz domains were deforming at different times under heterogeneous strain conditions.

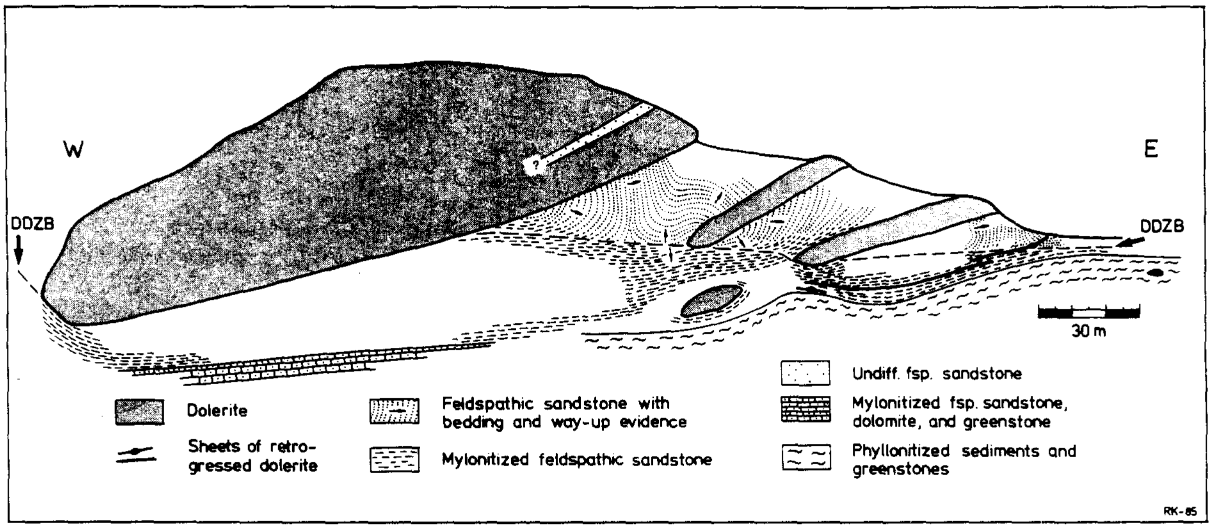


Fig. 10. Profile containing the upper ductile deformation zone boundary (DDZB) in a cliff-section at Ulvberget (located on Fig. 1). All the photomicrographs are from samples collected at Ulvberget except for Figs. 12(b) and (c).

The shear sense may also be discerned from parallel sets of shear planes which cut the mylonitic foliation at low angles and extend it in the direction of shear (Fig. 9d). These narrow shear planes, or C-planes, are zones of very fine-grained quartz and mica where slip is concentrated during progressive shear. The quartz ribbons form little intrafolial folds confined to domains bound by the shear planes (Fig. 9d). Although the microfolding is locally heterogeneous due to the influence of the feldspar porphyroclasts, the overturned limbs generally record a sense of shear which agrees with the general SE transport direction displayed by the asymmetric augen structures and the quartz subgrain fabrics.

*Strain softening fabrics*

So far we have presented evidence that the Särv mylonites formed as a consequence of thrusting based on (1) the shear zone geometry of the dikes at the base of the thrust sheet, (2) the parallelism of the mineral stretching directions to the regional displacement direction of the Särv thrust and (3) the agreement between microstructural shear-sense indicators and the transport direction of the sheet. Mylonite formation must have been the mechanism for thrust propagation during a major portion of its movement history. Since there are no weak glide horizons within the surviving Tossåsfjället Group and the sheet was further reinforced by the Ottfjäll dolerites, it appears that strain softening may have been required to localize ductile flow in the fault zone. The very sharp strain gradient recorded by the decreasing dike-bedding angle towards the DDZ boundary is also shown by corresponding changes in the microstructures. We will briefly describe the salient features of the microstructural transition below, but the fabrics, grainsize reduction, metamorphic reactions, and deformation mechanisms are under investigation (Gilotti in prep.).

The progressive development of the Särv mylonites is illustrated at Ulvberget, a cliff section first figured by Strömberg (1955) and recently mapped by Kumpulainen (Fig. 10). The DDZ at Ulvberget is located in a locally overturned limb of a syncline at the southern edge of the Särv sheet (Fig. 1). Although the contact with the underlying Tännäs augen gneiss is not exposed here, the upper DDZ boundary is visible for a distance of over 300 m (Fig. 10).

Above the DDZ boundary, the sediments of the Lövan Formation are dominantly trough cross-bedded, medium to very coarse-grained feldspathic sandstones. The transition from unfoliated rocks to strongly foliated grey platy mylonites occurs over a distance of a few meters. The variable thickness of the DDZ boundary here is in part due to the presence of the large competent dikes and the primary range in grain size and composition in the Lövan Formation sediments.

The sandstones are composed of detrital quartz, feldspar and accessory hematite grains which lie in a fine-grained, metamorphic matrix of quartz and mica. Figure 11(a) shows the microstructure of a feldspathic sandstone sample located 10 m above the DDZ boundary at Ulvberget. Typically the protolith contains subangular clasts with small aspect ratios, lacks a throughgoing cleavage and has a low percentage of matrix material. Strings of opaque minerals mark the traces of cross-lamination and are approximately parallel to bedding. All of the grains show some amount of plastic deformation (Fig. 11b) which increases systematically towards the DDZ. Undulose extinction, core and mantle structure and deformation bands are common in the quartz grains; a number of grains have totally recrystallized to form aggregates of new grains. Stable fracturing (Mitra 1984) of feldspars has begun and was accompanied by the syntectonic breakdown of feldspar to quartz plus mica.

Progressive heterogeneous deformation of the sandstones led to grainsize reduction in the quartz–feldspar–mica mylonites via both plastic and diffusion–mass transfer processes. Dynamic recrystallization and recovery, the rotation of quartz clasts into favorable positions for intracrystalline glide, and the coalescence of clasts via pressure solution has produced extremely long (1–2 mm) and very narrow (0.05 mm) ribbons (Figs. 11c, 9b and c). The breakdown of feldspar to quartz plus mica was aided by the continual stable fracturing of feldspars so that the fine-grained ductile matrix fraction increased in volume (Fig. 11c). New lath-shaped micas are aligned along the ribbon boundaries (Fig. 9c) in the foliation planes and some of the new quartz has grown in the pressure shadows of the feldspars (Fig. 9a). Grainsize reduction may have led to further softening by changing the dominant deformation mechanisms from dislocation-controlled to grainsize-dependent processes (White *et al.* 1980, Schmid 1983). For example, displacement due to sliding is observed (Fig. 11c) in feldspar fragments which have slid past each other along fractures lined with rectangular micas. Grain-boundary sliding may also account for a large amount of displacement in the matrix in the shear planes where the mica laths are concentrated.

Syntectonic retrograde metamorphism of the dolerite dikes is a very important strain softening process in the Särv DDZ. Despite the variation in macroscopic appearance from dense aphanitic basalts with or without phenocrysts to medium-grained porphyritic rocks, the undeformed Ottfjäll dolerites are abyssal tholeiites of constant bulk composition (Solyom *et al.* 1979). The primary olivine tholeiite texture and mineral assemblage are well preserved in the Särv sheet except along the dike margins, which show varying amounts of alteration due to interactions with fluids from the sediments during intrusion (Claesson & Roddick 1983). At Ulvberget, olivine tholeiites with a sub-ophitic texture are preserved 30 m above the DDZ boundary (Fig. 12a). As the isotropic dikes were rotated and thinned into the DDZ, the olivine–clinopyroxene-Ca–plagioclase mineralogy was altered to an epidote + actinolite + biotite + oligoclase/albite (upper greenschist facies) assemblage.

The progressive change in the dolerite mineral assemblage is shown at Alsen (Fig. 1), where a single dike can be followed for over 50 m in the DDZ beneath a klippe of Särv Nappe lithologies (Gee & Kumpulainen 1980). Just above the DDZ, the least deformed dolerites contain relict phenocrysts of epidotized plagioclase and clinopyroxene mantled by blue–green actinolitic hornblende (Fig. 12b). The unfoliated matrix is composed of fine-grained epidote, plagioclase, blue–green amphibole and euhedral biotite. The armoured clinopyroxene persists to the DDZ boundary where it is totally replaced by amphibole; with increasing deformation these amphibole megacrysts were recrystallized into fine-grained aggregates. Inside the DDZ, the greenschist bands have a strong foliation defined by the alignment of the acicular amphiboles and neoblasts of brown biotite laths which, along with epidote/clinozoisite and oliogoclase/albite, form the fine-grained groundmass (Fig. 12c). The final product of the combined deformation and metamorphism of the dolerites is a greenschist mylonite composed of bands of epidote + biotite up to a few millimeters thick. Grainsize reduction of the dolerites was probably accomplished by diffusion–mass transfer processes (White & Knipe 1978) and aided by infiltration metasomatism (Beach 1980).

## CONCLUSIONS

The Särv thrust sheet is interpreted as a large horse with a classical ramp-flat geometry. Hangingwall anticlines are developed just inside the leading, lateral and oblique branch-lines; and the internal imbricate faults are minor structures. Except for the fact that the Särv thrust was never emergent, the structural geometry is similar in scale and style to the Pine Mountain block (Harris 1970) in the southern Appalachians. We believe that the structural geometry of the Särv thrust sheet also developed in the same simple manner, i.e. a general foreland progression of thrusts climbing up-section over three-dimensional footwall ramps. Late-stage megaboudinage (e.g. Gee 1978, 1982) is not required to produce the lensoid geometry.

Thrusting was accomplished via the development of a DDZ along the base of the sheet. A brittle–ductile type shear zone model (Ramsay 1980) describes the overall geometry of the basal Särv DDZ. This zone is brittle because of the juxtaposition of two different tectonic units (one with dikes over one without) and the hangingwall anticlines imply a major structural discontinuity; it is ductile because of the intense strain gradient recorded by the change in dike orientations and the development of mylonites in the fault zone. The lack of late-stage brittle features, such as a brittle fault surface or cataclasites parallel to the thrust zone, suggests that the Särv thrust was not active at shallow crustal levels (Sibson 1977) and supports the idea that the Särv sheet was further transported onto Baltica on lower thrust fault systems.

The shear-zone geometry combined with the microstructural evidence (stretching lineations and shear-sense indicators) supports the conclusion that the mylonites formed during thrusting. The lack of natural glide horizons in the Särv sheet requires some form of strain softening to create a weak zone for thrust motion. Once created, deformation was confined to the DDZ via continual softening processes dominated by metamorphic reactions in the epidote–amphibolite schists and aided by grainsize reduction in both the quartz–feldspar–mica and the dolerite-derived mylonites. An unknown amount of grain-boundary sliding probably accounts for the extreme attenuation of the dikes now represented by thin bands of greenschist mylonite.

The dike swarm reinforced the sheet so that it behaved as a rigid block during thrusting and ductile flow was confined to a thin, strain-softened mylonitic layer (the DDZ) at the base of the sheet. In a qualitative way, the deformation profile given by the dikes (e.g. Fig. 2) records an extreme strain gradient which is compatible with a material that is deforming pseudoplastically, as Schmid (1975) suggested for the Glarus thrust. This situation is fundamentally different from the Hubbert & Rubey (1959) model, because rather than sliding on a fracture with a high fluid pressure, the Särv sheet was continuously deforming in a zone of confined flow at its base.

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