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An investigation of lower crustal deformation: Evidence for channel flow and its implications for tectonics and structural studies

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Abstract

High-grade metamorphic terranes generally have similar deformation fabrics characterised by a shallowly dipping transposition foliation and recumbent folds. Deformation paths are typically non-coaxial and strain magnitudes extreme. We refer to this association of structures and metamorphism as the high-grade nappe association (HGNA), and argue that it is common, and represents crustal-scale (kilometres thick) shear zones. The structure of such rocks is generally interpreted as comprising stacked thrust sheets separated by thrusts. We concluded, however, that the 'thrusts', which are ubiquitously parallel to the transposition foliation, are not thrusts in the normal sense, but are various discontinuities, rotated by large pervasive shear strains. The thrust-like appearance may be further enhanced by late localised shear strain.

The sense of shear may be constant across an HGNA shear zone, giving it the geometry and kinematics of a crustal-scale (kilometres thick) detachment zone. Alternatively it may reverse across the body, consistent with channel flow. Structural evidence therefore supports current ideas on the behaviour of the middle to lower crust during orogeny, and illustrates the deformation mechanisms involved. We describe the association with special reference to the Monashee complex, and discuss the implications of our interpretation for the kinematics of high-strain zones, palinspastic reconstruction, and interpretation of deformation fabrics at various scales.

Keywords: Shear zone; Channel flow; Strain; Vorticity; Kinematics; Nappe

1. Introduction

High-grade regional metamorphic rocks, in our experience, generally have a horizontal or shallowly dipping composite foliation (see also Fyson, 1971 and references therein; Mattauer, 1973) unless the foliation is reoriented by later structures. Where reoriented, the composite foliation generally has an enveloping surface that is shallowly dipping, suggesting that the dip of the foliation was initially shallow. Steeply dipping high-grade shear zones are an exception, but they are mostly flanked by shallowly dipping high-grade rocks, and are rarely voluminous compared with the latter. These shallowly dipping rocks have similar fabrics. They are characterised structurally by their prominent composite foliation, abundant intrafolial folds, and various kinematic indicators, which indicate that the foliation developed by transposition during a non-coaxial deformation. They are the infrastructure rocks of Wegmann (1935).

On a large scale, sequences of such rocks may be divisible into individual nappes, recognisable by differences in lithostratigraphy or metamorphic grade, or by the presence of faults or localised high-strain zones (mylonites and phyllonites), between different groups of rocks. Such discontinuities are generally parallel to the transposition foliation. The similarities of fabric lead us to group these rocks into a high-grade structural association, which we refer to here as the high-grade nappe association (HGNA). We believe that many high-grade regional metamorphic rocks belong to this association.

The discontinuities in rocks of the HGNA are commonly interpreted as thrusts (e.g. Gee, 1975a,b; Coward, 1980; Boullier and Quenardel, 1981; McClay and Coward, 1981; Boyer and Elliott, 1982; Butler, 1982, 1983 with discussion by Duncan, 1984; Platt, 1984; McDonough and Parrish, 1991; Brown et al., 1992; Gibson et al., 1999), the analogy

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being drawn with brittle thrusting as exemplified by the Rocky Mountain thrust and fold belt. Rocky Mountain thrust and fold belt terms such as 'sole thrust', 'roof thrust' and 'duplex' have been adopted to describe the structure of some high-grade areas (e.g. in the southern Canadian Cordillera, Brown et al., 1986; Brown and Journeay, 1987; McNicoll and Brown, 1995; Crowley, 1999; Gibson et al., 1999). In HGNA rocks of the Canadian Cordillera, a thrust, the Monashee décollement, has been proposed (Brown, 1980; Read and Brown, 1981; Brown and Read, 1983; Brown et al., 1986; Journeay, 1986), and interpreted as a sole thrust toward which all Rocky Mountain thrusts converge (Brown et al., 1992; Cook et al., 1992). This implies that the deformation mechanisms and the associated styles and geometries of structures of the Rocky Mountain thrust and fold belt are applicable to rocks of the HGNA. Based on geometrical, mechanical and field constraints, we argue here that this interpretation is fundamentally in error. Deformation at the mid- to deep-crustal level is characterised primarily by penetrative flow, not by stacking of discrete thrust sheets.

A number of writers have related nappe formation to ductile thrusting, separating the nappes by the shear zones, or even allowing internal deformation of the nappes by shear (e.g. Henderson, 1981; Krabbendam et al., 1997; Gibson et al., 1999; Treagus, 1999; Rose and Harris, 2000). We explore this approach further and argue that the shear strain is important and the whole HGNA fabric evolved in crustal-scale shear zones. Shear strains are large, but the thrust-like discontinuities are not mechanisms of deformation, and are not even necessarily the locus of above-average shear. They are inherited and, in general, involve far less thrust-like displacement than what is suggested by the mismatch across the discontinuity. We correlate the HGNA with the broad zones of detachment and channel flow being explored by modellers (Davidson et al., 1997; Beaumont et al., 2001a,b; Jamieson et al., 2002).

After characterising the association, we interpret the structural development and tectonic setting, and then discuss some of the implications of our interpretation. We make special reference to the Monashee complex of the Canadian Cordilleran Omineca Belt (Gabrielse et al., 1991), as a typical example of the association. This complex comprises two structural highs, the northern Frenchman Cap culmination and the southern Thor–Odin culmination. Our work is concentrated in Thor–Odin.

2. The high-grade nappe association

2.1. General description

As outlined above, the defining characteristics of the HGNA are: high-grade metamorphism, the horizontal to shallow dipping transposition foliation (or its enveloping surface), a non-coaxial deformation history, and where divisable into sub units, sheet-like bodies (individual nappes) with boundaries that are parallel to the transposition foliation. This broad description fails to cover many details of the structure as well as the variations that occur from one area to another. We discuss this more fully here and note that despite variation, there is an amazing similarity between areas.

The transposition foliation is defined by compositional layering, which is typically of mixed origin. Some layers may represent primary compositional variation such as sedimentary layers, plutons and lava flows. Others represent dykes and veins that have been intruded at various times prior to the end of transposition. All are in various stages of being folded into asymmetrical intrafolial folds (e.g. Fig. 1), as a result of the progressive deformation. The rocks are commonly migmatitic, and leucosome is also generally folded and rotated into the transposition foliation. Orthogneisses are generally tabular, parallel to the transposition foliation and intensely deformed, indicating that their shape is secondary and a product of the deformation.

Folds in the HGNA are usually abundant and vary greatly in scale (e.g. Fig. 2). Microscopic to outcrop-scale folds are generally present, and larger folds including regional-scale structures occur in some areas (e.g. Fig. 2), but are absent or difficult to recognise in others. For example, in the Monashee complex the large folds have amplitudes up to 20 km (Figs. 3 and 4), and various authors (e.g. Thomas, 1979; Treagus, 1999) show folds in the Tay Nappe of Scotland with amplitudes of the order of 40–50 km. At Little Broken Hill, Australia, folds with amplitudes measurable in hundreds of metres have sheath-fold geometry (Williams, 1967) as do regional folds in the Foxe fold belt of northern Canada (Henderson, 1981). The fold hingelines are commonly approximately parallel to a stretching lineation, or in the case of sheath folds, the lineation parallels the axis of the sheath.

HGNA fabrics are those typical of shear zones—environments of large finite strain accumulated non-coaxially. There is commonly a grain-shape fabric inclined to the transposition layering at angles of the order of 30°. This is interpreted as an *S*-foliation (commonly a steady-state foliation, cf. Means, 1981), in normal shear-zone usage (Berthé et al., 1979), whereas the transposed layering (S_T) is approximately parallel to the shear plane (*C*-foliation; Berthé et al., 1979). Locally, where folds are incompletely transposed, there may be an axial plane foliation, which is inclined to both the *S*-foliation and the transposition foliation (S_T). Kinematic indicators such as *C*/*S* relationships, rotated porphyroblasts/porphyroclasts, fold asymmetry, shear bands (C' of Berthé et al., 1979), and rotated boudins all indicate a non-coaxial deformation history and the same sense of vorticity through large parts of, if not throughout, an entire zone. For example in the Tay Nappe the sense of rotation of garnets neither changes across the axial planes of folds (Williams et al., 2000), nor across the full thickness of the nappe (cf. Krabbendam et al., 1997; Rose and Harris, 2000). In the Thor–Odin culmination of the

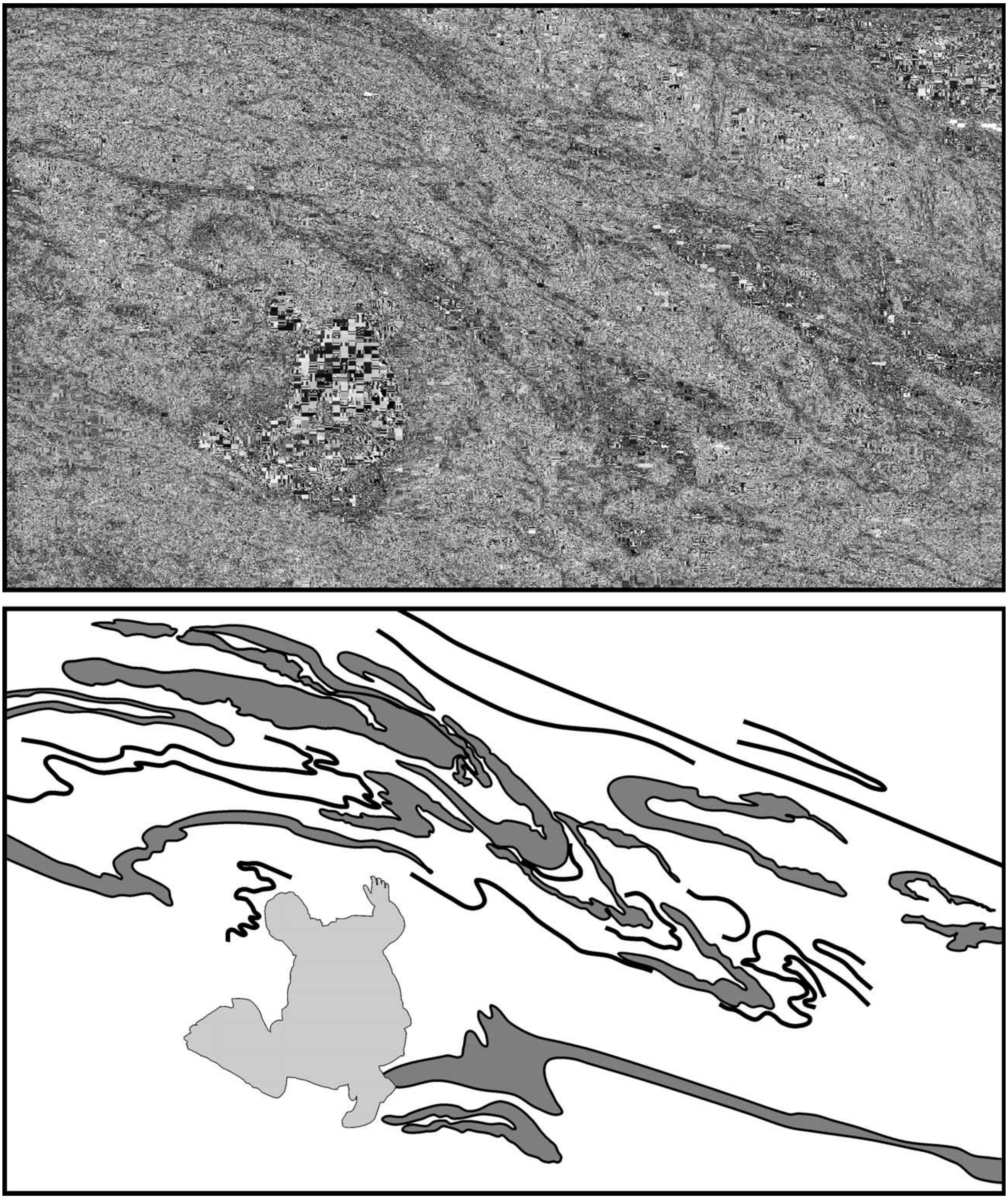


Fig. 1. Example of transposed migmatised gneisses and schists from the Thor–Odin culmination of the Monashee complex, Canadian Cordillera. Larger folds outlined in the tracing below. The folds shown here are F_1 and F_2 and the transposition foliation is present at all scales. It is defined by the alignment of the fold limbs and discontinuities at all scales; the microscopic foliation, locally visible, is defined by transposed chevrons and crenulations. Person for scale.

Monashee complex (Fig. 3) the asymmetry of mesoscopic folds is constant throughout most of the sequence, as is the asymmetry of coeval C–S–C' fabrics (Jones, 1959; Johnston et al., 2000; Spark, 2001).

The discontinuities that separate individual nappes may be faults or high-strain zones. The latter may be mylonitic or phyllonitic, and may be of similar metamorphic grade to the nappes (coeval with the nappe fabric), or may be retrograde

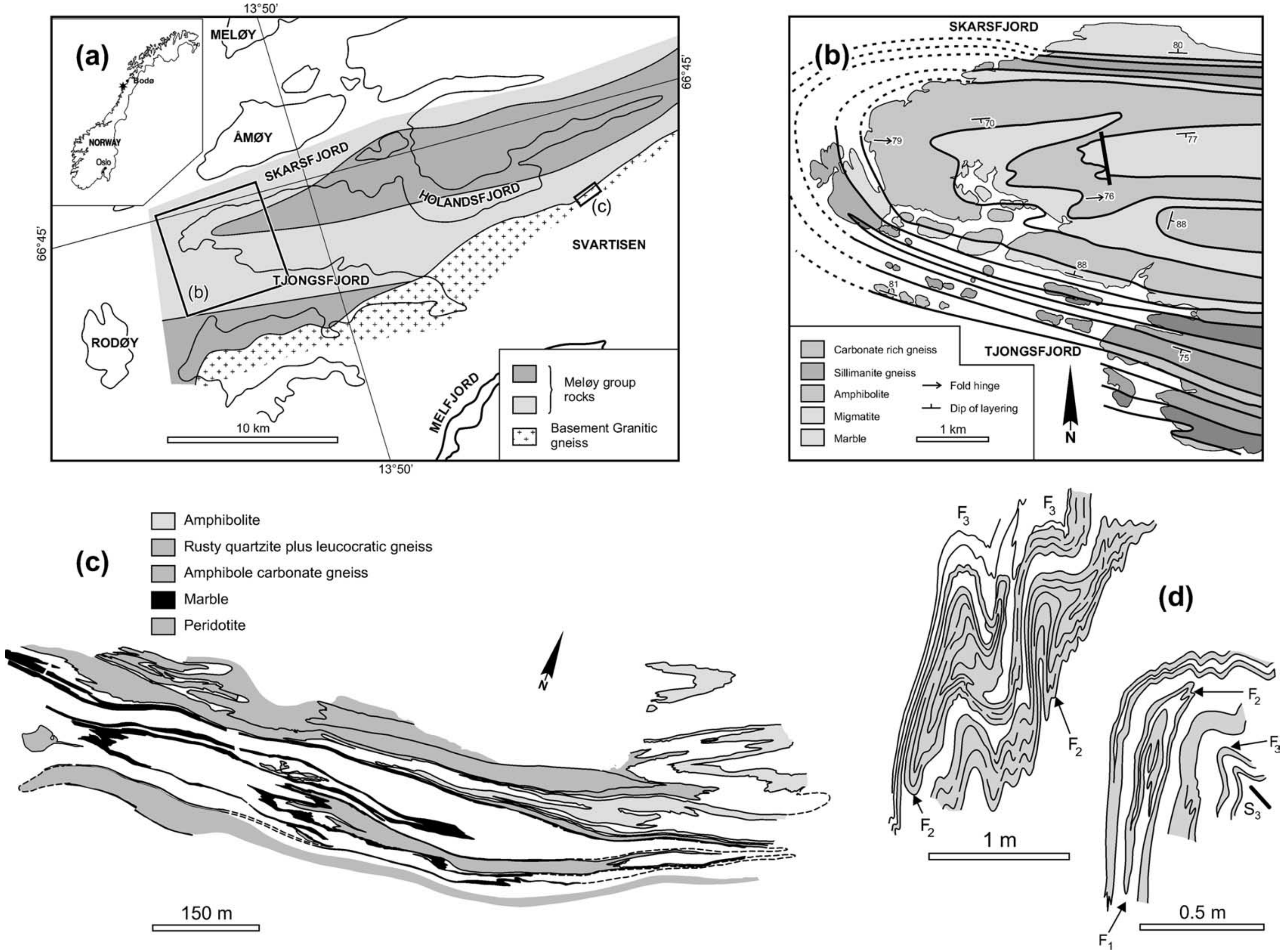
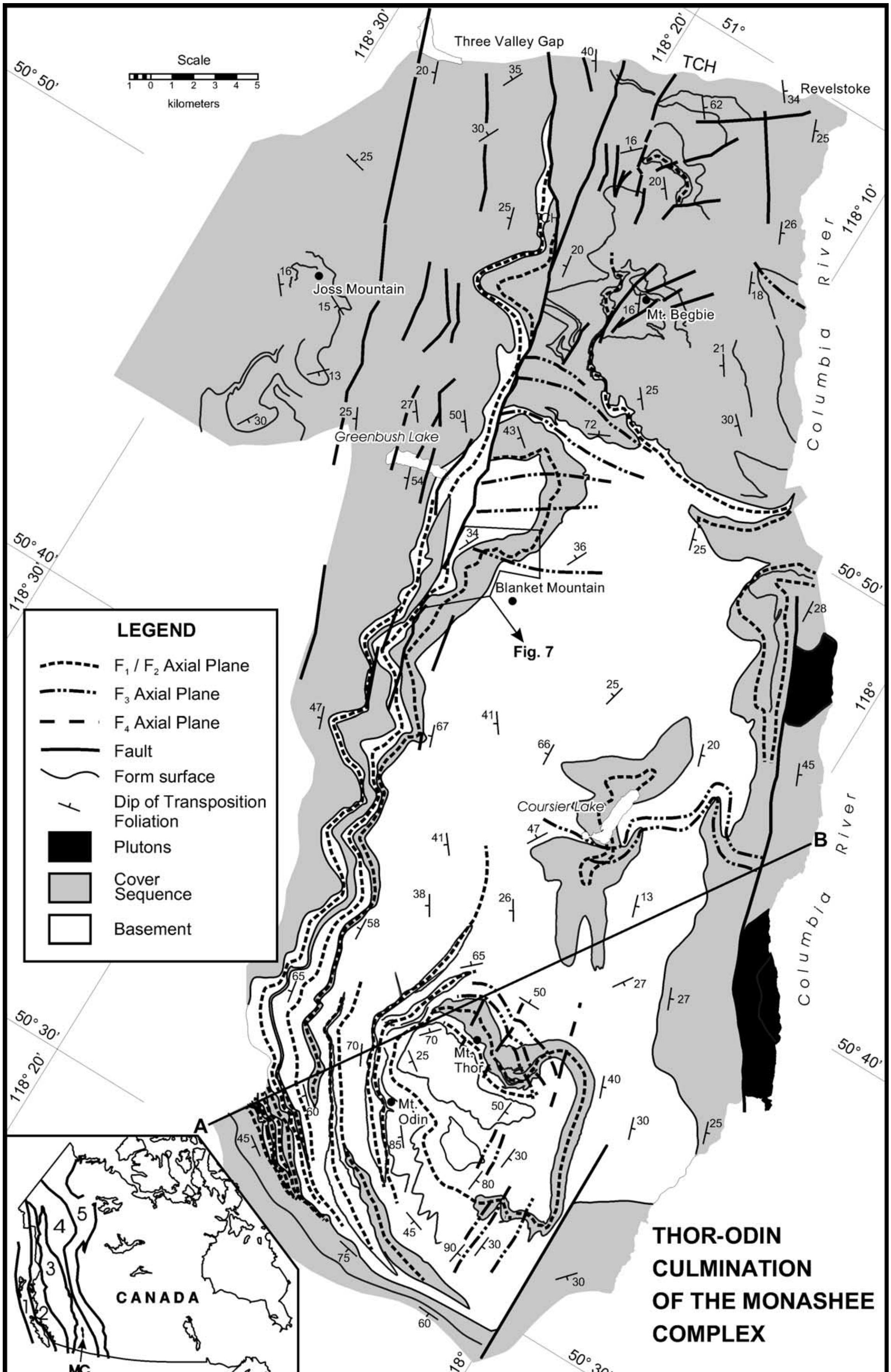


Fig. 2. Transposition in the Svartisen area, central Norway (location indicated by star south of Bodø on the inset map), showing variation in the scale of folds related to transposition. (a) Regional synformal structure of Holandsfjord area after Rutland and Nicholson (1965). Locations of maps ((b) and (c)) are indicated. (b) Detailed map of peninsula between Skarsfjord and Tjongsfjord showing two generations (F_2 and F_3) of tight regional scale folds. Based on mapping of Kuipers (1982). (c) Detailed map (after Williams, 1983) of large mesoscopic folds at the head of Holandsfjord below the Engabreen. Dips of fold limbs are steep to vertical and folds plunge moderately towards west. (d) Small mesoscopic folds (after Williams, 1983) from the Holandsfjord area.


Scale
1 0 1 2 3 4 5
kilometers
Three Valley Gap
TCH
Revelstoke
Joss Mountain
Mt. Begbie
Greenbush Lake
Blanket Mountain
Fig. 7
Coursier Lake
Mt. Thor
Mt. Odin
Columbia River
A
B
LEGEND
F1 / F2 Axial Plane
F3 Axial Plane
F4 Axial Plane
Fault
Form surface
Dip of Transposition Foliation
Plutons
Cover Sequence
Basement
CANADA
MC
THOR-ODIN CULMINATION OF THE MONASHEE COMPLEX

(post-nappe fabric). They may contain all the same structures and show the same deformation history, differing only in being more strongly transposed. For example in the Swedish Caledonides (Williams and Zwart, 1977) folds in the nappes have hingelines that define a girdle that lies in the transposition foliation. The more open folds tend to be perpendicular to a mineral lineation (interpreted as a stretching direction) and tighter (more evolved) folds are progressively closer in orientation to the mineral lineation, the tightest being approximately parallel. In the high-strain zones between the nappes, fold morphologies and distribution are qualitatively the same. Quantitatively, however, tighter, and locally sheath folds, parallel or nearly parallel to the lineation are the most common, indicating a greater magnitude of strain than in the nappes themselves.

The nappes generally have an extremely thin aspect ratio, being truly sheet-like. For example in the Caledonides of central Scandinavia the nappe pile, comprising at least four major nappes and several smaller ones, is generally only 3 km thick (maximum thickness 7 km), but covers an area approximately 250 by 250 km. These nappes are recognisable because of variation in lithology, stratigraphy and metamorphic grade, and have high-strain zones between them (Gee, 1975a,b). Elsewhere, through-going discontinuities are less abundant. For example, in the Thor–Odin culmination of the Monashee complex, rocks belonging to the association are more than 10 km thick, but there are no recognisable regional discontinuities (Reesor and Moore, 1971; Johnston et al., 2000; Spark, 2001; Kruse et al., 2004) other than late brittle faults. Regional-scale recumbent isoclinal folds (Figs. 3 and 4; see also Reesor and Moore, 1971) were interpreted by Duncan (1984) as nappes, separated by thrusts. Truncation of structures (Fig. 5) does occur and a lack of continuity in the folded sequence indicates that discontinuities are present, but none are regionally persistent and they are themselves folded locally. They are sharp with the appearance of brittle faults but are completely healed (Fig. 5), i.e. there are no fractures preserved. Exceptional continuity of outcrop makes it possible to recognise the discontinuities as asymmetrical boudinage and shear surfaces following S_T and C', at various scales, rather than discrete nappe-scale faults or shear zones.

We restrict the association to high-grade metamorphic rocks, including amphibolite facies, high-pressure equivalents and anything of higher temperature. A similar structural association may occur in some lower-grade rocks (for example the Morcles Nappe of the Alps (e.g. Casey and Dietrich, 1997) and the Köli Nappe of Sweden (Gee, 1975a,b)), but at lower grades continuity is commonly better preserved and upright folds and foliations are generally the norm. In HGNA rocks, metamorphism is typically late relative to the transposition. For example, near Blanket Mountain in Thor–Odin (Fig. 3), kyanite crystals overprint early folds and are folded by late folds, both of which are associated with transposition, and are then pseudomorphed by sillimanite, which was still stable during a later crustal-scale extensional deformation (cf. Johnston et al., 2000). Elsewhere in the Monashee complex, assemblages that we associate with the transposition have been shown to represent pressures at least as high as 8–10 kb and temperatures around 750–800 °C (Norlander et al., 2002). They are overprinted by lower pressure (<5 kb) assemblages at approximately the same temperature (750–800 °C) (Norlander et al., 2002), which we associate with the onset of crustal extension. Microstructural observations indicate that high-temperature conditions outlasted ductile deformation and there was abundant leucosome development throughout the transposition and during the ductile phase of the crustal extension.

As in the Swedish example above, sequences of rocks belonging to the HGNA are commonly several kilometres thick and kilometres to tens or even hundreds of kilometres in aerial extent. In our experience most regional high-grade rocks belong to this association (cf. Fyson, 1971 and references therein). They therefore represent significant volumes of the Earth's crust. Examples with which we are familiar include the Caledonian Nappes of central Scandinavia (e.g. Trouw, 1973; Williams and Zwart, 1977; Williams, 1983), the Trans-Hudson–Orogen Kisseynew domain of Manitoba and Saskatchewan (Lucas et al., 1994; Norman et al., 1995; Kraus and Williams, 1999), the Teslin Zone of the Yukon (de Keijzer et al., 1999), the Thor–Odin culmination of the Monashee complex (Reesor and Moore, 1971; Johnston et al., 2000), the Little Broken Hill area of NSW, Australia (Williams, 1967) and the Sesia–Lanzo of the Italian Alps (Williams and Compagnoni, 1983). In this paper we are concerned primarily with the high-temperature/medium-pressure as opposed to high-temperature/ high-pressure members of the association (to which the Sesia-Lanzo example belongs).

2.2. Typical deformational history of the HGNA: an example from the Monashee Complex

The deformation history of HGNA rocks in general, as indicted by overprinting of structures, is monotonously similar. It involves progressive deformation with several generations of folds associated with the transposition (e.g. Central Scandinavia, Little Broken Hill, the Trans-Hudson Orogen and the Monashee complex; see references above). We briefly describe the transposition folds of the Thor–Odin culmination of the Monashee complex as a typical example of the HGNA.

Fig. 3. Map of the Thor–Odin culmination of the Monashee complex of the Omineca Belt, Canadian Cordillera (based on Kruse et al., 2004). The inset map shows the location of the whole Monashee complex (MC), comprising Frenchman Cap to the north and Thor–Odin to the south. Only Thor–Odin is represented in the detailed map. Zones delineated and numbered 1–5 on the inset map represent the Insular Belt, Coast Belt, Intermontane Belt, Omineca Belt and Foreland Belt of the Canadian Cordillera, respectively. TCH: Trans-Canada Highway. Section line A–B (Fig. 4) and source area of data presented in Fig. 7 are indicated.

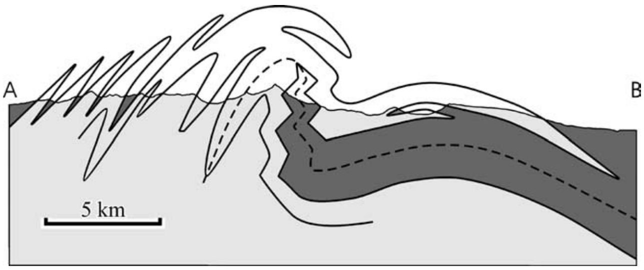


Fig. 4. Interpretive section through the south end of Thor–Odin (for position of section see A–B in Fig. 3). Dark shading represents the cover and light shading the basement. The dashed line represents the axial surface trace of a regional F_1/F_2 fold, refolded by a regional F_3 in the centre of the section. The overall antiformal appearance is due to a north-trending upright regional F_4 fold. The section passes rapidly from the F_3 hinge, which it cuts obliquely, into the steep F_4 limb (see Fig. 3). This makes the regional F_3 fold appear almost upright, whereas it actually has an axial surface dipping shallowly to the SW. To the west the rocks pass into an F_4 synform and to the east they pass into the Columbia River Fault. No vertical exaggeration.

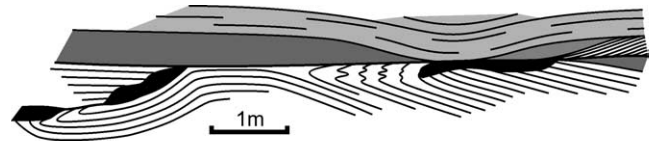


Fig. 5. Discontinuities in quartzite (dark grey), laminated feldspathic quartzite (grey) and calc-silicate (form surface only) rocks from the Thor–Odin culmination of the Monashee complex. Black areas represent pegmatite. Outcrop viewed looking west. Note the truncated F_1/F_2 fold in the calc-silicate unit. The only evidence of the discontinuities is the observed truncations of fold and foliations.

Rocks of the Monashee complex can be divided into core and cover sequences, the base of the cover commonly being marked by the base of a thick quartzite. The boundary between the two is interpreted as an unconformity (e.g. Fyles, 1970; Høy and McMillan, 1979; Journeay, 1986; Ross and Parrish, 1991) and it is, therefore, significant that the transposition foliation is everywhere parallel to the unconformity, both above and below. Previously the area has been divided into an eastern para-authochthon and a western allochthon (the Selkirk Allochthon) separated by the Monashee décollement (Brown, 1980; Read and Brown, 1981; Brown and Read, 1983; Brown et al., 1986; Journeay, 1986; Gibson et al., 1999; Crowley et al., 2001). However, extensive mapping across the proposed décollement in Thor–Odin (cf. Kruse et al., 2004) has failed to reveal any structural discontinuity other than late brittle faults (Johnston et al., 2000; Spark, 2001). The transition from Monashee complex to overlying Selkirk Allochthon along the west side of Thor–Odin is complicated by the faults (Fig. 3; Kruse et al., 2004), but in the south where the sequence is continuous, rocks of the Selkirk Allochthon and the Monashee complex cover are inter-folded, with the same transposition fabric present in both, and a gradual change from Monashee complex below to Selkirk Allochthon above. One argument for the décollement is the presence on Joss Mountain (Fig. 3) of Devonian intrusives (Johnston et al., 2000), which are found elsewhere in the Selkirk Allochthon, but not in the Monashee complex (e.g. Woodsworth et al., 1991, p. 495) suggesting that the Selkirk Allochthon is allochthonous. However, in view of the large amplitude folds (Fig. 4) and the normal faults separating Joss Mountain rocks from the Monashee complex, it is not necessary to propose a décollement to explain how they could have been widely separated in the Devonian. Thus, we interpret the whole of the Monashee complex, and that part of the Selkirk Allochthon covered by the map (Fig. 3), as a single crustal-scale zone of transposition. It is still useful to distinguish rocks west of the normal fault passing through Greenbush Lake from those to the east of the fault. We therefore refer to them as the Joss Mountain complex, east of the fault being the Monashee complex.

In Thor–Odin, three generations of folds are recognised (Fig. 6). F_1 and F_2 folds are tight to isoclinal, generally dismembered and oriented with their axial surfaces subparallel to the transposition foliation and their hinges typically parallel or acutely inclined to a stretching lineation, which forms a partial girdle with its maximum in the south-west quadrant (Fig. 7c). They commonly occur as asymmetrical fold pairs and occur in all sizes up to regional scale structures (Figs. 1, 3, 4, 6 and 8). They may have a foliation (the transposition foliation) parallel to their axial plane or may be overprinted by a shape foliation. The only sure way of distinguishing F_1 and F_2 folds is by overprinting. Where they can be distinguished, the F_2 folds (Fig. 6) tend to be more open than F_1 and their axial surfaces and hinges more inclined to the local transposition foliation and stretching lineation, respectively. Map-scale F_1/F_2 folds appear to have few metre-scale parasitic folds. Small folds do occur on their limbs but the asymmetry is mostly the same on both limbs, consistent with interpretation of the large folds as early (see below) and the small folds as younger dragfolds.

F_3 folds (Figs. 6, 8 and 9) occur at all scales with amplitudes up to 2 km. They deform the transposition foliation and generally lack an axial plane foliation. Where an axial plane foliation is present, it is inclined to the transposition foliation and generally restricted to symmetrical folds (Fig. 8c), or to the hinges and common limb of asymmetrical fold pairs. It is not normally developed on long regional fold limbs, and even where best developed, is patchy (Fig. 8c). When present it is generally a crenulation cleavage or is defined by transposed micro-chevrons, and is progressively rotated into the transposition foliation. F_3 folds are typically more open than F_2 folds (Figs. 6, 8 and 9), but there is a gradation between the two, both in orientation and in style. The tight F_3 folds are distinguishable from F_2 folds only by virtue of their continuity with more open F_3 folds or by overprinting relationships. F_3 folds lie on the same partial girdle as defined by F_1 and F_2 folds, but spread between the normal to the stretching lineation and a westerly orientation, which overlaps the spread of the stretching lineation (Fig. 7b).

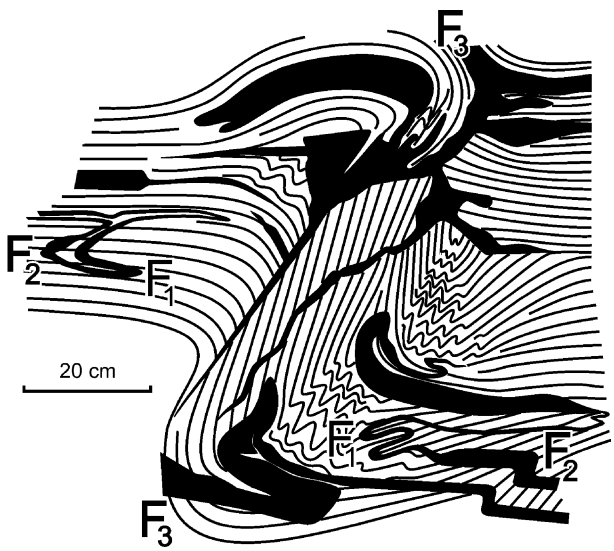


Fig. 6. Overprinting between F_1, F_2 and F_3 folds in quartz–feldspar–biotite migmatitic gneiss, from the Thor–Odin culmination of the Monashee complex. Black areas represent leucosome of various ages. Outcrop viewed looking NW. F_3 fold is seen in profile, other folds seen in oblique sections.

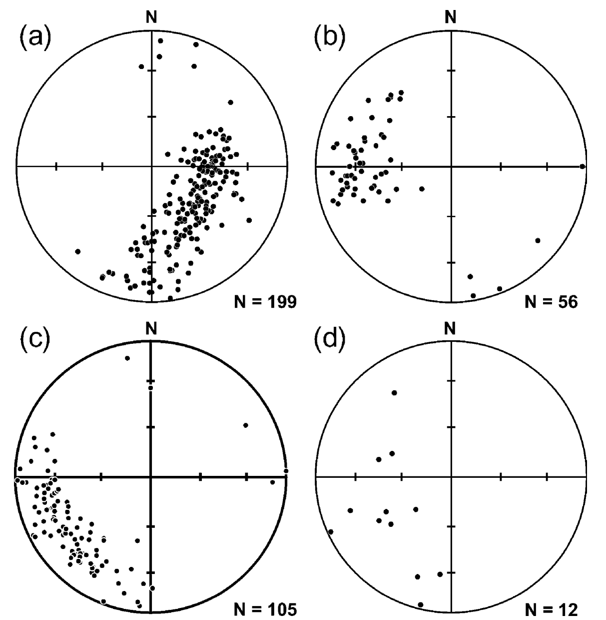


Fig. 7. Orientation data (equal area projection) for area immediately NW of Blanket Mt (see Fig. 3) of the Thor–Odin culmination of the Monashee complex. (a) Poles to the transposition foliation forming a girdle about a macroscopic F_3 fold. (b) Plunge of mesoscopic F_3 folds; these folds developed as dragfolds about an approximately west-northwest-trending vorticity vector and have started to rotate towards a south-southwesterly plunging stretching direction. (c) Stretching lineation spreading between the original top-to-the-north north-northeasterly flow direction and a late (D_5) west-southwesterly crustal extensional direction. (d) F_1 and F_2 folds in various stages of rotation towards the SSW plunging stretching direction. These folds have z-asymmetry if viewed looking west-northwest. The stretching lineation is interpreted as developing with a south-southwesterly plunge parallel to the early, top-to-the-north-northeasterly flow. Metre-scale F_1–F_3 folds mostly developed as dragfolds parallel to the vorticity vector, and therefore, initially plunged west-northwesterly. The folds rotated initially towards the stretching lineation, but then locally the lineation and folds were rotated towards the late (D_5) crustal extension direction. Thus the stretching lineation spreads on a girdle between the early and late extension directions, but since positive identification of folds is difficult where D_5 is well developed, there are no fold measurements in such areas. The plots of F_1–F_3 folds therefore do not reflect the D_5 extension.

Regional F_3 folds in Thor–Odin (Fig. 9) have long limbs mostly dipping shallowly in the south-west quadrant, with rare parasitic F_3 structures and shorter limbs with abundant parasitic F_3 folds and an enveloping surface that dips steeply in the north-east quadrant. The axial surfaces of these parasitic F_3 folds vary considerably in dip, mostly between vertical, through southwesterly dipping, to shallowly northeasterly dipping. The steep dips are most common in the rare parasitic folds on the shallowly dipping regional fold limbs and recumbent attitudes are most common on the steep regional limbs (Figs. 8 and 9).

In most of Thor–Odin, asymmetrical F_3 folds verge predominantly north-northeasterly (Jones, 1959; Johnston et al., 2000). The vergence of F_1/F_2 folds is variable, due to their rotation towards the stretching lineation. However, taking into account sense of asymmetry and orientation, they are consistent with the same vergence, and rare F_1/F_2 sheath folds also verge toward the north-northeast. In the Joss Mountain complex the same styles and generations of folds are recognised as in the Monashee complex and the vergence of F_1/F_2 is the same. F_3 vergence, however, changes to southwesterly.

The boundaries of this crustal-scale shear zone do not occur in Thor–Odin or in the Joss Mountain complex. However, the upper boundary is recognisable in the work of Campbell (1970) and Murphy (1987), at a higher structural level, in the Selkirk Allochthon of the neighbouring Cariboo Mountains. They report upright folds that become progressively recumbent downwards, the structural transition coinciding with a metamorphic transition from greenschist to amphibolite facies. A similar situation is reported by Zwart (1979), from the Pyrenees, at the boundary between the low grade superstructure and high-grade infrastructure rocks. Upright folds in the superstructure deflect into recumbent folds of the infrastructure and the latter has the characteristics of the HGNA.

From the preceding description it is obvious that strain magnitudes are large in the Monashee complex and the Joss Mountain complex, but no accurate quantification is possible. However, an impression of the minimum magnitude can be obtained. For example, in a simple-shear[1] progressive deformation, periclinal folds develop approximately parallel to the vorticity axis with axial planes approximately perpendicular to the flow direction (Fig. 10).

[1] It is generally sufficient to consider simple shear only, both here and below. The existence of a pure-shear component only changes the amount of strain required to rotate pre-existing planes or lines to certain angles with respect to the shear plane and direction, not the resulting geometry.

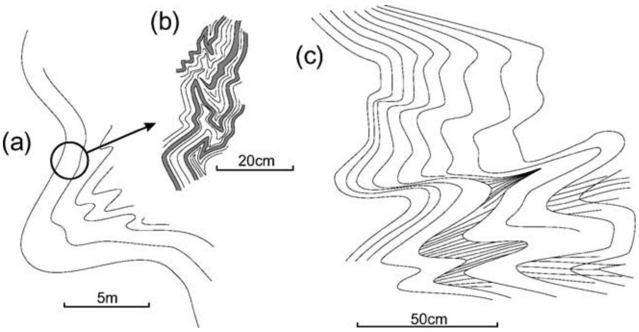


Fig. 8. F_3 mesoscopic folds on the steep limb of a regional F_3 fold in the Blanket Mt area (see Fig. 3). Both viewed towards the west. (a) and (b) Overprinting of F_1 or F_2 folds by F_3. (c) F_3 folds showing variable tightness and irregular development of an axial plane foliation.

For a fold plunging 10° (i.e. the fold axis is inclined to the assumed horizontal vorticity axis by 10°), a shear strain of 20 is required to rotate the fold axis to within 16° (a reasonable estimate of the angle commonly observed) of the projection of the shear direction on the axial plane (cf. Skjernaa, 1989). The closer the orientation of the initial fold axis to the vorticity axis, the larger the shear strain required to rotate it a given amount (Fig. 10b).

Such a shear strain is compatible with the observation that foliations on both sides of the unconformity in the Monashee complex are parallel. Even if the initial situation was one of an angular difference of up to 90° across the unconformity (Fig. 11a), a shear strain >10 will render the difference too small to recognise in the field (Fig. 11d), for most orientations of the original markers. However, even a shear strain of 20, in simple shear (maximum principal stretch of 20.2), is far too small to explain commonly observed isolated metre scale boudins in large outcrops (several km^3). So there is evidence of even larger strains and we are confident, therefore, that a shear strain of 20 is a minimum value, within the limitations of the simplifying assumptions. Similar estimates can be made for the other examples of the HGNA listed above. For example in the outcrops represented in Fig. 2c, isoclinal F_2 folds (with amplitudes an order greater than their wavelength) are parallel to the lineation and can be traced through a series of boudins (i.e. both limbs occur in each boudin) where the lengths of the boudins and distances between them are equal.

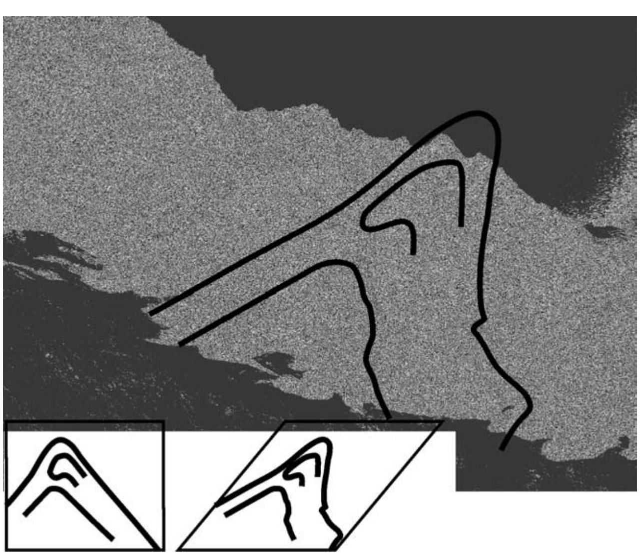


Fig. 9. F_3 fold on the east side of Mt Thor (see Fig. 3) overprinting an earlier F_1 or F_2 fold, viewed towards the west. Visible amplitude of F_3 fold is estimated at 70 m. Inset drawings show interpretative history of folds. An early upright dragfold, overprinting an earlier isocline, is modified by NE directed simple shear into an overturned structure, with a younger buckle fold on its steep shortened limb.

2.3. Timing of deformation in the Monashee complex

There has been extensive dating in the Monashee complex (e.g. Carr, 1992; Parrish, 1995 and references therein; Crowley and Parrish, 1999; Gibson et al., 1999; Crowley et al., 2001), and U–Pb dates primarily from zircon and monazite, are generally interpreted as representing peak metamorphism and deformation. The dates young downwards from ~170 Ma at a high structural level in the Selkirk Allochthon, to ~78 Ma at the highest level in the Monashee complex, to ~49 Ma in the core of the Monashee complex. At the deepest levels of the Monashee complex, in the Frenchman Cap culmination, the rocks are regarded as having experienced essentially no deformation since the Palaeoproterozoic (Crowley, 1999), despite the lack of any obvious change in fabric, and apparently no metamorphic peak temperature high enough to reset zircon, monazite and titanite. The downward younging of peak metamorphism is explained by progressive burial by thrust sheets advancing from the west (Parrish, 1995). The Monashee complex, being the lowest sheet, was the last to be buried (Parrish, 1995; Crowley and Parrish, 1999; Gibson et al., 1999; Crowley et al., 2001). However, the lack of regional scale thrusts does not support this interpretation.

We suggest a different interpretation that is consistent with the structural observations, involving detachment and channel flow in warm middle to lower crustal rocks, sandwiched between strong upper crust and strong Palaeoproterozoic basement. Since deformation is outlasted by peak metamorphism it may have started considerably earlier than the peak so that part of the history may not be represented by the dates (cf. Kuiper, 2003). Initially a broad detachment zone developed and evolved into a zone of channel flow. At this stage the upper boundary may have been migrating upwards (cf. Kuiper, 2003). The channel tunnelled through cooler rocks (cf. Jamieson et al., 2002, for example), which became involved in deformation as they warmed. This, combined with cooling from above due to progressive unroofing, resulted in both channel boundaries migrating downwards. Finally, conditions reverted to detachment flow which by then involved basement rocks, but ended within the basement, where warm and cool basement rocks were juxtaposed (Crowley, 1999). We

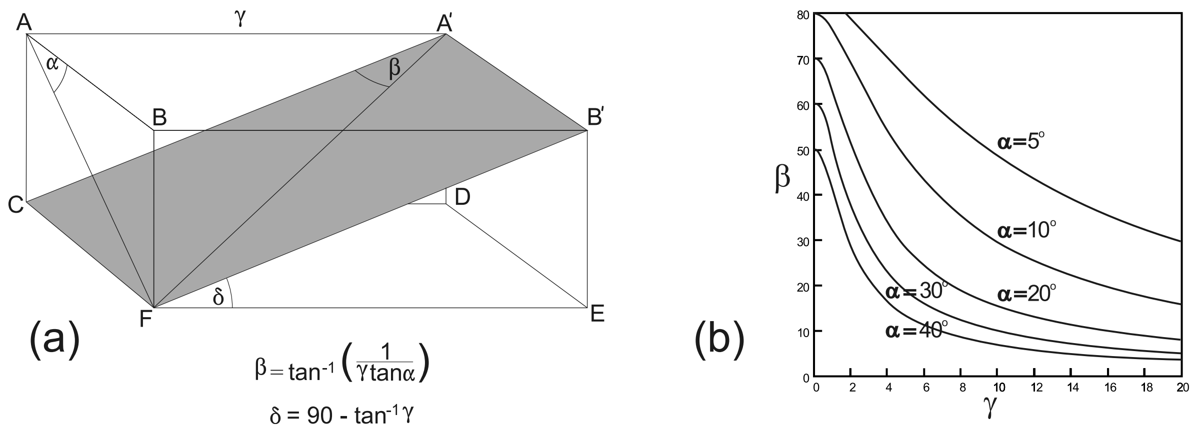


Fig. 10. (a) Relationship in simple shear, between the initial plunge (α) of an upright fold with axis AF and axial plane ABCF and the final orientation of the passively rotated axis and axial plane, as a function of shear strain (γ). The shear plane is AA′BB′ and the shear direction AA′. (b) Graph of the angle (β) between the rotated fold axis (A′F) and the projection of the shear direction on the axial plane of the fold, as a function of initial plunge (α) and shear strain (γ).

suggest that this is the sort of complexity that might be expected in channel flow zones in general.

3. Development of the HGNA

3.1. Introduction

The salient deformational features that occur in the HGNA as a group (not necessarily recognised in all examples), which have to be explained, are as follows: features found in all examples include (1) a crustal-scale shear zone characterised by a transposition fabric resulting from folding before and/or during progressive non-coaxial deformation, (2) either a constant vergence across the full width of exposed rocks or a simple reversal, and (3) high-grade metamorphism that generally continued late relative to the deformation. Features only found in some examples, (1) regional folds that are so large that they cannot be explained by any simple process of buckling or bending in response to shortening (there is insufficient room in the crust (cf. Fig. 4), and (2) regional thrust-like discontinuities.

Because the association is commonly interpreted in terms of the Rocky Mountain thrust and fold belt type thrusting model, we first discuss the inadequacies of this model as applied to the fabric of HGNA rocks, and then present an alternative model that is capable of explaining the observations.

3.2. Inadequacies of the foreland thrust and fold belt model when applied to the HGNA

Although nappes separated by discontinuities can resemble the structure of a foreland thrust and fold belt, with the nappes resembling the thrust sheets and the discontinuities the thrust faults, there are fundamental differences between the two structural associations aside from metamorphic grade. The main difference is that Rocky Mountain thrust and fold belt type thrust sheets are only weakly deformed, so that continuity of internal primary structure is preserved, whereas primary structure in the HGNA is completely transposed, indicating intense internal deformation. In the Rocky Mountain thrust and fold belt situation, deformation is generally brittle, and is essentially concentrated on the thrust plane. In the HGNA, deformation is penetrative and generally (but not exclusively) ductile,

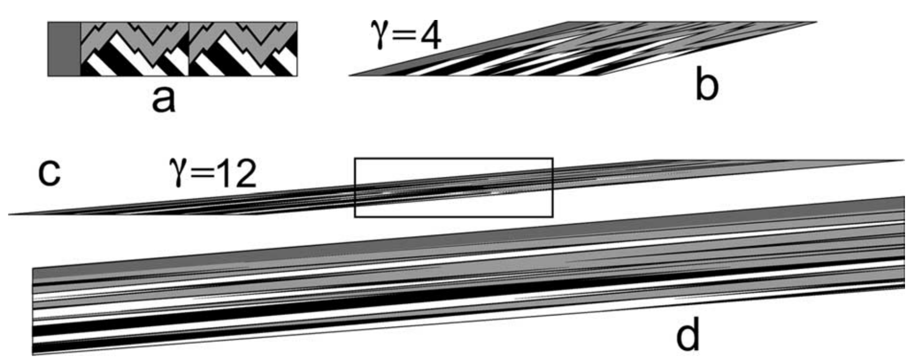


Fig. 11. Modification of upright folds and an angular unconformity by simple shear. At a shear strain as low as 12, all markers are inclined to one another at sufficiently small angles as to make it difficult to recognise non-parallelism in the field. (d) Details of the rectangular area indicated in (c); it contains several fold hinges as well as the unconformity.

and there is not necessarily any recognisable variation in the intensity of deformation across nappe boundaries. Folds are not a penetrative feature of the Rocky Mountain thrust and fold belt structure at outcrop- or smaller-scale. Intrafolial folds, however, are characteristic of the HGNA, at outcrop- to micro-scale and commonly at regional scale also. It follows that, although the aspect ratio of Rocky Mountain thrust sheets may have changed, they have always been sheet-like with their long boundaries parallel to original primary fabric. On the other hand, the original shape of the HGNA nappes is unknown because of intense folding and strain.

The Rocky Mountain thrust and fold belt model has been applied to the Monashee complex in Thor–Odin for example (McNicoll and Brown, 1995), but despite almost continuous outcrop there are no mappable thrusts separating their individual horses, nor any fault–bend folds. There are, however, large isoclinal folds that do not fit the interpretation. Boyer and Elliott (1982) applied the Rocky Mountain thrust and fold belt model to the Alpine Nappes, but their interpretation is refuted by Casey and Dietrich (1997). Similarly, Ramsay (1997) has criticised model-driven application of Rocky Mountain thrust and fold belt type thrusting in general, and with particular reference to development of the ductile fabric of the area of the Moine Thrust in NW Scotland.

The major problem with applying the Rocky Mountain thrusting model to high-grade rocks is, therefore, that the model offers no explanation of the intense deformation observed within the nappes. In the Rocky Mountain situation, the thrusts and related structures, including folds, can all be explained as products of thrusting and there is a sound mechanical basis for the process (e.g. Elliott, 1976). However, in the HGNA, the fabric indicates that all rocks were deforming penetratively during the deformation process, and therefore changing their aspect ratio and shape. We are unable to envisage a mechanically sound process, which involves ductile thrusting, of internally flowing thin sheets, over one another.

3.3. Alternative interpretation of the HGNA

All features indicate that rocks of the HGNA have formed in regional-scale (typically kilometres thick and tens to hundreds of kilometres in extent) horizontal to shallowly dipping shear zones and we interpret the various structures in the context of this general deformation environment. We start by considering what might be expected to happen to various pre-existing structures if subjected to such deformation (cf. Krabbendam et al., 1997; Williams, 1999).

Fig. 12 shows three simple combinations of folds and faults before and after applying varying amounts of simple shear. In each example, pre-existing discontinuities that are initially inclined to layering are rotated until almost parallel to layering. Irrespective of their origin (e.g. normal or transcurrent fault) they develop thrust-like geometry as all markers are rotated towards parallelism with the shear plane. With a suitable initial geometry, old layers may be placed on young (Fig. 12 a and b), so that after a large shear strain, the whole sequence has the appearance of a stack of thrust sheets (Fig. 12c). At a shear strain of 20 all original features are so nearly parallel to one another that any discrepancy would be difficult to recognise in the field. Fault-drag folds, in a horizontal shear zone, become tight

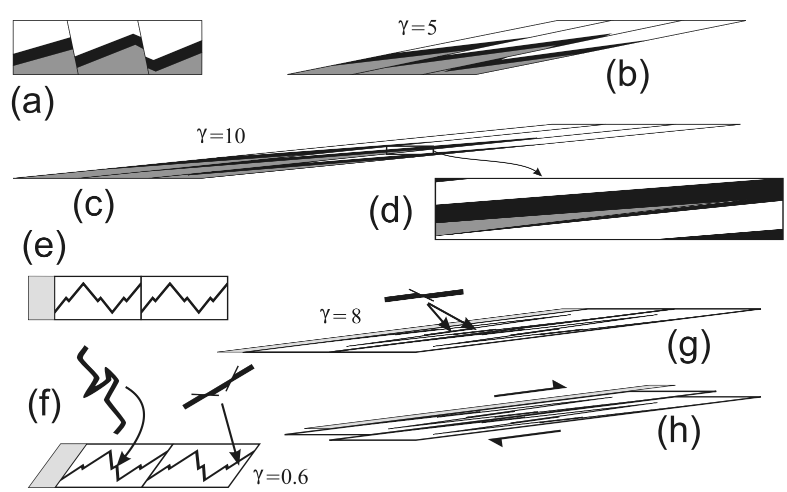


Fig. 12. Modification of early structures by large magnitude simple shear. (a)–(d) Early normal faults (a) are converted progressively (b) to nappe-like structures (c). Initial 'fault-drag' (a) is converted to tight recumbent folds with attenuated limbs between alternate fold-pairs ((c) and (d)). Pre-existing upright folds (e) are converted to recumbent folds and intervening discontinuities again develop thrust geometry (g). Initial sub-equant markers, such as the shaded rectangle at the LH end of (e), are converted to thin tabular bodies (g); this is the expected fate of granitoids believed to be the protolith of tabular granitic gneisses. Alternate limbs of the upright folds (e) undergo continuous stretching or initial shortening followed by stretching. This may lead to secondary structures such as buckle folds (cf. Fig. 9), boudinage and shear bands (f). Late shearing parallel to the early thrust-like discontinuities and/or parallel to S_T may produce faults, mylonites or shear zones with thrust geometry (h) and apparent thrust-related fabrics. These features are expected to form as S_T approaches parallelism with the shear plane, but are shown earlier here because of the problem of representing larger shear strains.

recumbent folds with alternate limbs apparently attenuated (Fig. 12d), and pre-existing upright folds (Fig. 12e) are rotated towards the shear plane to become recumbent isoclines (Fig. 12g). One limb is in the extensional field throughout shearing, but the other shortens initially. Consequently, parasitic buckling and an axial plane foliation can be expected on the latter (Figs. 9 and 12f). Once the transposition foliation is approximately parallel to the shear plane, local perturbations can amplify into dragfolds (cf. Bons, 1993), which then rotate and contribute to the transposition. Fold hinges rotate progressively towards the shear direction.

The interpretation of folds as presented here differs significantly from interpretations based on the slip fold model (Weiss, 1959; Ramsay, 1960) in which the folds are a product of shear-zone-parallel shear (e.g. Casey and Dietrich, 1997; Gibson et al., 1999). At the strain magnitudes believed to pertain in many examples of the association considered here, such folds would reverse in vergence and then unfold (e.g. Ramsay et al., 1983). The inherited folds and dragfolds of our interpretation rotate towards the shear plane but never truly reach it unless overprinted and rotated by other structures such as later dragfolds. Thus, they become tighter and more attenuated but maintain the same vergence. Another difference between the two models is that the one presented here does not require changes in shear sense across every axial plane as is required by the slip fold model. Our interpretation is thus consistent with the observation that the shear sense is constant across many axial planes, through thick packages of transposed rock.

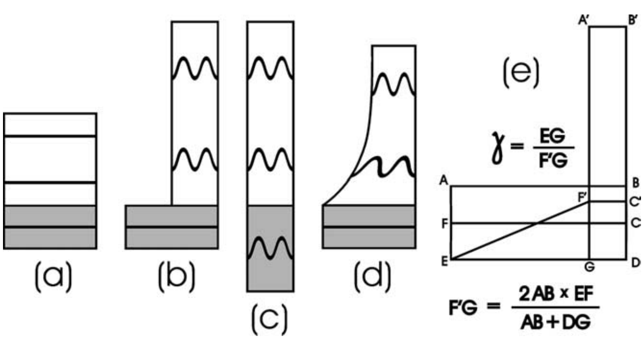


Fig. 13. End-member behaviour during shortening of the crust (white) over upper mantle or strong crust (grey). This represents for example, what is happening to the crust and mantle in Fig. 15b. (a) Initial configuration. (b) The situation assuming a frictionless boundary between crust and strong substrate. The crust deforms in bulk pure shear by folding and potentially reverse faulting. (c) and (d) The situation assuming that the crust is attached to mantle. In (c) both weak crust and substrate deform and in the end member they have the same strength and deform equally. In (d) the substrate is rigid and the weak crust deforms in bulk pure shear near the surface, but has to undergo a non-coaxial deformation at depth in order to maintain continuity with the substrate. Early folds develop as upright structures, but locally they are modified by shear to form overturned folds. (e) Simplified representation of initial and final form of the weak crust, according to the end member represented in (d). See text for further discussion.

The gradation in style and orientation as exemplified for example by the Monashee complex of Thor–Odin indicates that F_1–F_3 folds are consistent with a progressive deformation in a flat-lying shear zone. A simplistic interpretation of the Monashee complex is that F_1 folds were inherited from earlier deformation and modified by shearing (Figs. 12e–g and 13) to produce a transposition foliation (S_T). S_T was then perturbed by F_2 dragfolds that, on tightening, further contributed to the development of S_T. Finally, F_3 folds developed and deformation stopped before their limbs had been completely transposed back into S_T. All hingelines began to rotate as they developed, as described above. The most rotated are the F_1/F_2 folds and the least rotated are F_3. Thus, F_2/F_3 represent a progressive deformation in which early folds have completed the cycle of perturbation–transposition and late folds have been frozen in the perturbation stage. In reality, many F_1 folds may be dragfolds, but in view of the fact that individual folds can be traced across the upper boundary of the shear zone, in the Omineca Belt (Murphy, 1987) and in the Pyrenees (Zwart, 1979) at least some F_1 folds must be inherited. They may be much older or may represent an early stage of the deformation that resulted in development of the shear zone (see Fig. 13). It is also possible that more than one generation of folds is inherited.

Two types of HGNA shear zone have been described, one with constant sense of shear and another across which the sense of shear reverses. We cannot rule out the possibility that the latter is the norm and the former simply an artefact of inadequate exposure. For example, one side of the shear zone may be eroded or the other may be too deep to be exposed. However, it seems likely that there is a progression from one type to the other (see Discussion below) so that we might expect both to exist. Consequently, the non-coaxial, high-strain structural characteristics and the high-grade of metamorphism lead us to interpret the association in terms of (1) detachment flow within the crust and/or at the base of the crust, where the sense of shear is constant, and (2) channel flow (e.g. Beaumont et al., 2001a,b), where the sense of shear reverses across the channel.

4. Discussion

4.1. Discontinuities

We now discuss the nature and significance of discontinuities that are commonly interpreted as thrusts, in the HGNA.

Despite detailed mapping in the Monashee complex, we have been unable to find through-going thrust-like discontinuities. Discontinuities occur at various scales but there are no regionally persistent surfaces that dissect the complex into thrust sheets or horses. It is possible, however, to arbitrarily divide the complex into various nappes defined by individual regional recumbent folds (cf. Fig. 3). If it were not for the continuity of glacier-polished outcrop, one might imagine that these folds were separated by through-going discontinuities since whole segments of various units are truncated. However, because of the exceptional outcrop we are able to see that discontinuities, though numerous, are local. It may be that this is a more common situation than realised, because few regions are as well exposed. Of those listed above that we are familiar with, only the Svartisen area is comparable and it also lacks through-going discontinuities in the area mapped.

We have shown that where thrust-like discontinuities do exist, they may be a product of the interaction of the strain and pre-existing discontinuities. This implies that such structures have existed throughout the shear-zone deformation, and we would therefore expect that they would themselves be folded in areas such as the Monashee complex, where large dragfolds are present. There is, however, the possibility that faults develop during the shear-zone deformation as large shear-band type structures or if the shear zone is widening (in the sense of Means (1995)), they might develop as reverse faults. Either would ultimately be transposed into a thrust-like structure. Some faults or high-strain zones may develop parallel to the shear plane.

Some discontinuities have mylonite fabrics consistent with their development parallel to the shear plane, but many are ambiguous. However, if a mylonite is retrograde for example, it cannot have been inherited and preserved through high-grade metamorphism. Nevertheless, we suggest that these are not evidence of thrusting in the normal sense, but rather indicate reactivation of an earlier surface or zone that has been rotated into approximate parallelism with the shear plane (Fig. 12h). The early features may be, for example, inherited faults, shear zones, lithological contacts, or weak lithological units that have acted as planes or zones of weakness. In shear-zone terminology these would be *C*- or *Y*-shears. The shear strain associated with the mylonite might be small, but because of the pre-existing discontinuity the structure would have the appearance of a major thrust. The mylonite below the Glarus nappe of the Swiss Alps may be an example of such a situation.

Schmid (1975) concluded that an unreasonable strain rate would have been necessary for the mylonite below the Glarus nappe to accommodate nappe emplacement, given the known constraints. He discussed various possible solutions to the problem, including the possibility that shear deformation was not restricted to the narrow mylonite, but occurred across a broader zone. An alternative possibility is that the Glarus thrust is a rotated earlier discontinuity and the mylonite is due to later reactivation. The geometry described, with cleavage inclined to the discontinuity and with folds parallel to the stretching lineation close to the discontinuity and perpendicular to the lineation further away, is consistent with the HGNA, despite the lower (greenschist) metamorphic grade.

In view of all the possibilities we believe that there is no basis for interpreting discontinuities between nappes as thrusts. This has implications for palinspastic reconstructions as discussed below.

4.2. Implications for timing of deformation

The concept of diachroneity of fold generations is not new (e.g. Hobbs et al., 1976, p. 355; Williams and Zwart, 1977) and diachroneity is likely to be acute in shear zones where local rates of deformation may vary considerably and where the boundaries of the zone may be migrating, progressively or intermittently, into or out of the adjacent country rock. We find it useful to use the designations F_1, F_2, etc. where the generations are identified by overprinting, but stress that F_1 in one outcrop can be younger than F_2, or even F_3, in another (Williams, 1985). The folds have developed progressively and F_1 designation implies maturity of development as opposed to a less mature F_2. Where shear-zone boundaries migrate through time and activity varies in rate from place to place within the shear zone, there will be no simple relationship between maturity and absolute age.

In environments such as Benioff zones it is conceivable that the same non-coaxial flow may be generated repeatedly in different packages of rocks. Thus, adjacent packages may be deformed at very different times, but the kinematic picture may be the same for each, and thus, the fabric is the same in appearance and orientation. Similarly, if horizontal zones of ductile detachment or channel flow occur deep in orogens in general, as suggested by the abundance of HGNA rocks, it is possible that a detachment or channel flow fabric generated by one mountain building episode could conceivably be reactivated by a later one. If the whole zone was reactivated, evidence of the earlier flow might be annihilated. However, if only part of the zone was reactivated it is possible that fabrics of very different age would be geometrically indistinguishable. A further complication might occur where, as a channel approaches the surface, plug-like flow might develop due to extreme localization of deformation into channel margins, with a central portion of the channel being carried by the surrounding flow without internal deformation.

4.3. Implications for kinematic vorticity determination in high-strain zones

Our observation that the compositional layering in HGNA is a transposition foliation has implications for the kinematic vorticity determination in shear zones. All vorticity number determination methods, besides other assumptions (e.g. Jiang and White 1995), rely on the existence and recognition of a flow extensional eigenvector (fabric attractor of Passchier (1997)). Compositional layering in shear zones is commonly assumed to represent the flow extensional eigenvector, but since the layering is generally a transposition foliation (S_T) this is unjustified. In the case of a rotating S_T toward the shear zone boundary orientation, the total vorticity with respect to the shear zone boundary is the sum of the vorticity relative to S_T plus the vorticity accommodated by the spin of the layer with respect to the zone boundary. Therefore, the kinematic vorticity number determined from the rock fabric assuming that S_T is the fabric attractor, might be expected to be significantly less than the bulk kinematic vorticity number. This may explain the low kinematic vorticity numbers consistently obtained by many authors (Simpson and De Paor, 1997; Xypolias and Koukouvelas, 2001; Bailey et al., 2004; Grasemann et al., 1999; Law et al., 2004). These authors have interpreted the low kinematic vorticity numbers as indicating large values of pure shear component in natural shear zones. We regard these kinematic vorticity numbers as not being representative of the bulk kinematics of the zone.

4.4. Implications for palinspastic reconstructions

If 'thrusts' in HGNA rocks do not necessarily have significant movement on them, there are important consequences for palinspastic reconstruction. A simple comparison between our interpretation and the thrust model is made in Fig. 14, where homogeneous simple shear is assumed across the zone. The difference between the original width and the final width of the zone is greater for the normal klippe-to-fenster (e.g. Hobbs et al., 1976, p. 309) method of reconstruction than it is for the flow model presented here (compare Fig. 14c (ii) with c (iii)). Pre-nappe-formation shortening is ignored in both models, even though early folding is probably a pre-requisite of the flow model. Unfortunately, we have no way of addressing this problem, other than to note that in the flow model a shortening in excess of 0.5 is probably required for development of the folds.

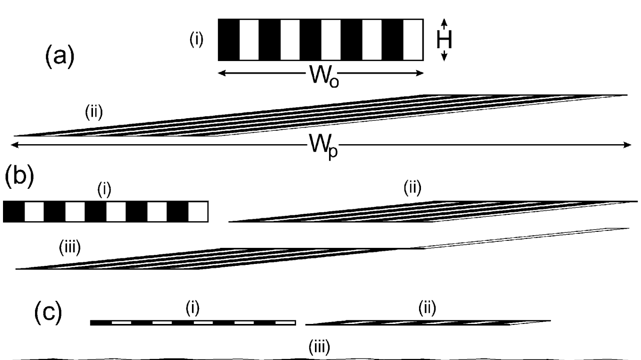


Fig. 14. Palinspastic reconstruction: comparison of penetrative flow versus thrust-stacking origin of nappes. Each part shows a section through a block of crust divided by discontinuities into 10 units with nappe-like geometry (ii). According to the flow model proposed here, the present geometry resulted from shear modification of the original geometry. (i) Is obtained by reversing the shear. (i) Is thus the restoration of (ii) according to the flow model. Restoration of the section according to the thrust-stacking model is achieved by reversing the thrusts individually (iii), according to the klippen-to-fenster method (i.e. each nappe is placed end to end with its neighbour. Only one thrust is restored in (b (iii)) because of space limitations; all are restored in (c (iii)). Aspect ratios and shear strains are as follows: (a) AR=5, $\gamma=10$. (b) AR=10, $\gamma=10$. (c) AR=50, $\gamma=10$. The original width of the deformed zone (W_O), for the flow model, is given by $W_O=W_P-\gamma H$ (where W_P is the final width of the deformed zone), and for the thrust model, by $W_O=(N^2H^2+(\gamma NH-\gamma H+W_P)^2)^{1/2}$. Comparison of (a)–(c) indicates the effect of aspect ratio on the reconstruction according to the thrust model.

The klippe-to-fenster method always gives the larger value for the original width of the zone (Fig. 14), but all other values being equal, the difference between the two models decreases as the aspect ratio increases. For a given aspect ratio, increasing γ or N (Fig. 14) also increases the difference between the predictions of the two models. The predictions of the klippe-to-fenster model can exceed those of the flow model by an order of magnitude, but for wide zones combined with small values of γ and N, the difference can be negligible. However, γ is difficult to determine and unless it can be shown to be small, there will be considerable uncertainty in palinspastic reconstruction, in addition to that which generally exists, due to imperfect knowledge of the geometry of the orogenic zone.

For example, a 200 km (W_P) wide zone with a thickness (H) of 10 km, combined with 10 nappes and $\gamma=10$ (Fig. 14b (ii)), according to the thrust model, gives an initial width of 1105 km, compared with an initial width of 100 km predicted by the flow model. Even if a pre-flow shortening of 75% is assumed for the flow model, the initial width would only be 400 km. Increasing the shear to 20 almost doubles the initial width predicted by the thrust model, so that $W_O=2102$ km as opposed to the more realistic estimate of 400 km.

Another problem for palinspastic reconstruction is one of provenance. If the thrusts associated with nappes are in fact inherited discontinuities, it becomes important from the point of view of reconstructing provenance to know their early history. For example, if a 'thrust' is really a rotated transcurrent fault, any interpretation based on the assumption that it is a thrust is misleading. There may have been considerable mass movement perpendicular to the direction of detachment or channel flow. In addition, a history involving varying amounts of detachment flow and channel flow, which are likely to be difficult to quantify, makes interpretation of provenance difficult, even if there is no movement perpendicular to the flow direction.

4.5. Tectonic implications

The HGNA is common and it is argued that regional high-grade metamorphic rocks in general belong to the association. High-pressure members of the association probably develop in Benioff zones (cf. Fyson, 1971; Mattauer et al., 1981) where the prime requirements of non-coaxial flow, large strains and high-grade metamorphism exist. Most members, however, are too hot (relative to pressure) to develop in Benioff zones and we suggest that they are evidence of the subhorizontal flow of the middle to

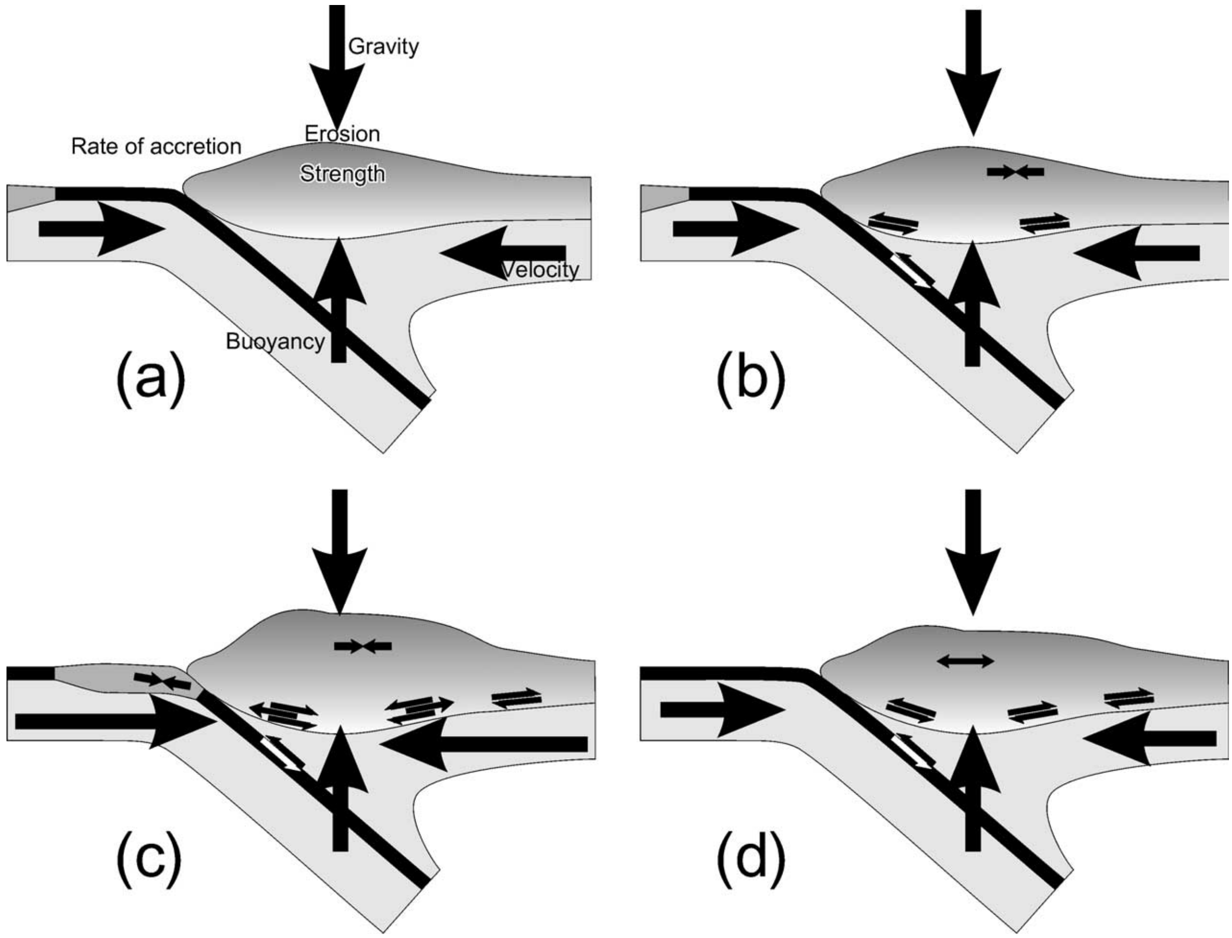


Fig. 15. Tectonic model for the development of detachment flow and channel flow. (a) Factors influencing the geometry of the orogen. (b) Tectonic erosion or downward flow of the mantel, and localised thickening of the crust result in detachment shear at the base of the crust. (c) An increase in convergence rate and input of crustal material result in further thickening of the crust and increased thermal weakening of its base. This results in the evolution of detachment flow into channel flow. (d) A decrease in convergence rate and reduction in input of crustal material results in collapse of the orogen and channel flow reverts to detachment flow.

lower crust that has been postulated by modellers (e.g. Beaumont et al., 2001a,b). We believe that the structural evidence in the HGNA overwhelmingly indicates that the middle to lower crust is generally weak on the orogen scale, and that it is a locus of flow within the lithosphere. This does not preclude the possibility of the detachment zone extending down to, or even into the mantle, or of the existence of other zones, but it does indicate that part of the crust is generally involved.

Mountain chains are a product of crustal shortening, and geological evidence indicates that the shortening is achieved at shallow levels by upright folding and reverse faulting, generally under greenschist facies or lower conditions. As we go deeper into middle to lower crust amphibolite facies conditions, we generally move into a regime of shallowly dipping fabrics (cf. Fyson, 1971; Mattauer, 1973, p. 185). This is consistent with shear resulting from shortening of the crust over strong basement crust or mantle (Fig. 15b). At shallow levels shortening is by a bulk-pure-shear process, involving upright folding and reverse faulting, and at deep levels there is an increasing influence of the comparatively strong mantle. Fig. 13 shows three end member situations. In (b) the boundary between the crust and mantle is frictionless and the whole crust deforms in bulk pure shear. In (c) the mantle is no stronger than the crust and both shorten in pure shear. In (d) there is perfect coupling between the crust and mantle, and the mantle is rigid. If any end member applies it is likely to be (d), but more likely is some combination of the three. In this detachment situation, the shear will be localised in any rocks that are significantly weaker than average. Thus, it could be concentrated anywhere in the crust rather than immediately above the mantle (cf. Jamieson et al., 2002). Typically, the weak rocks will be high-temperature and/or wet rocks, but we submit that the shear must occur whatever the conditions, so long as there is shortening of the crust. This may explain why strains and strain paths typical of the HGNA are found in some low-grade rocks (e.g. the Morcles and Köli nappes, see above).

If the contrast in strength is sufficiently high and the crust is thick enough (Fig. 15c), detachment flow may progress into a form of extrusive flow, or channel flow (Beaumont et al., 2001a,b), as evidenced by the reversal of shear in the Himalaya (Burchfiel et al., 1992; Grujic et al., 1996) and the Monashee complex. The fact that the vergence of folds in the Monashee only changes in the late stages of shear deformation (i.e. during F_3, but not during F_1 or F_2) is consistent with this interpretation. It is significant in this regard that high temperatures, high pressure, and development and emplacement of melt are synchronous with transposition (see above). Subsequently, a slowing down of the convergence rate or a reduction in the amount of material being accreted will result in collapse and thinning of the crust and channel flow might be expected ultimately to revert to detachment flow (Fig. 15d) driven by the extension.

This interpretation is consistent with the modelling of Beaumont et al. (2001a). The detachment stage is represented, for example, by $t=7.5$ My in fig. 4a of Beaumont et al. (2001a) where the shear sense is the same across the full width of the high-strain zone (F_1/F_2 in the Monashee complex and Selkirk Allochthon). As the slab thickens, the shear zone develops into a zone of channel flow (e.g. $t=52.5$ My in fig. 5a of Beaumont et al., 2001a) so that locally the sense of shear reverses (F_3 in the upper Monashee complex and the Selkirk Allochthon).

A potential test of the detachment model is to check observed minimum-strain magnitudes against model predictions. Fig. 13e diagrammatically shows the detachment model in the form that maximises shear strain. For (1) a shortening of 60%, which combined with erosion would keep the crust to a reasonable thickness, (2) a final detachment zone thickness of 10 km, as estimated for the Monashee complex, and (3) a shear strain of 42 at E (Fig. 13e) decreasing linearly to 0 at D (Fig. 13e) to give comparable values to those recorded in the Monashee complex; the length of the zone (ED) is 700 km. If we assume that E represents the approximate centre of the mountain chain, the full width of the chain is $\sim$1400 km. Further, the minimum shear strain suggested for the Monashee complex ($\gamma=20$) would only exist out as far as 350 km from the centre. Compared with observations, the width of the orogen is large, the pure shear shortening is reasonable, but the shear is minimal, suggesting that for the model to be viable, there has to be some other factor such as tectonic erosion of the mantle at the Benioff zone. Once channel flow develops there is not the same dependency of shear-strain magnitude on the width of the mountain chain and the pure-shear shortening.

As modelled by Beaumont et al. (2001a,b), the channel flow is much better developed on the 'pro-side' (i.e. on the subducting plate side) than on the 'retro-side' of the orogen. However, there is probably sufficient flow on either side to explain the development of the HGNA. Further, by varying the boundary conditions, different variations on this theme are possible. Beaumont et al. (2001a,b) have shown, for example, the importance of erosion (see also Zeitler et al., 2001) in enhancing the flow. Also, in their model, shortening of the crust on the retro-side of the suture is superficial and does not involve shortening of the lower crust and mantle. This is a model constraint, and if it were to be relaxed, for example by allowing tectonic erosion of the mantle below the hanging-wall plate at the suture, channel flow could be greatly increased on the retro-side of the orogen.

The F_3 deformation of the Monashee complex and the Joss Mountain complex has been interpreted in terms of extrusive flow previously, which in terms of local kinematics can be identical to channel flow. However, as envisaged, the extrusion is a static process, in which a wedge of rock is simply squeezed out, without loss of continuity, deforming and thinning as it goes. The problem with this is that the driving force required to keep the thinning wedge moving has to increase rapidly (e.g. Jaeger, 1964, pp. 140–143). The channel flow model (Beaumont et al., 2001a,b) includes the bigger picture, and because of the thickening of the crust and continuous feeding of material into the channel; there is no need for thinning. In fact, the channel may thicken. Thus, the channel flow model is a dynamic model involving ongoing crustal shortening and it is believed to be a better representation of the general tectonic process.

From a metamorphic point of view both detachment and channel flow can explain the inverted metamorphic sequences commonly observed in orogenic regions (e.g. Spear, 1993; Jamieson et al., 1996). Detachment flow may transport warm rocks from the interior of the orogen over cooler rocks, towards its margins. In channel flow, warm rocks may 'tunnel' through cooler rocks (e.g. Jamieson et al., 2002). Significantly, inverted gradients are reported in the Monashee complex (Journeay, 1986) and in the Joss Mountain complex and Selkirk Allochthon gradients are normal (Campbell, 1970; Murphy, 1987; Johnston, 1998). In qualitative terms, therefore, temperature decreases symmetrically away from the centre of the proposed channel.

5. Conclusions

The HGNA is a product of crustal-scale shear zones, characterised by horizontal to shallowly dipping transposition foliation, non-coaxial deformation histories with a constant shear sense across much if not all of the zone, high-grade metamorphism, and in some examples regional-scale recumbent folds and/or nappes. It is common and rocks belonging to the association represent a significant volume of orogenic mountain belts. We suggest that many high-grade metamorphic rocks belong to the HGNA.

Nappes are generally explained in terms of thrust tectonics with thrusting on discontinuities or narrow shear zones. However, such an interpretation ignores much of the structural evidence (specifically the internal transposition fabric), which it is incapable of explaining.

The HGNA is best explained in terms of large-strain crustal-scale shear zones in which the axial planes of earlier (inherited and/or transposition related) upright folds, and various discontinuities, are rotated towards the shear plane. The folds are progressively amplified so that their ultimate amplitude is more closely related to shear-strain magnitude than to the initial folding process. Earlier discontinuities develop thrust geometry without the necessity of any displacement appropriate to thrusting. Thus the nappes and their related 'thrusts' are structures modified in the HGNA shear zones rather than the cause of the HGNA fabric.

Structural evidence indicates that crustal-scale shear zones not only occur as Benioff zones but are commonplace in lower pressure environments. This supports current ideas of channel flow. These shear zones start as detachment zones, and if there is sufficient thickening of the crust and rocks become weak enough, channel flow develops where the rocks are weakest. This typically coincides with migmatitic conditions. Channel flow may later revert to detachment flow, due to late collapse of the orogen.

In some realistic situations detachment flow is a necessity and, therefore, is not necessarily restricted to high-grade rocks. This explains the existence of fabrics similar to those described from the HGNA in Greenschist facies rocks.

Earlier discontinuities or lithological contacts, once rotated close to the orientation of the shear plane, may undergo *C*- (*Y*-) shear and thus give rise to late mylonites. In the situation where the discontinuity is a pre-shear fault, the mylonite is not likely to accurately reflect the total displacement that has occurred on the 'thrust'. It is unlikely, however, that the discrepancy will be recognisable in general.

It is conceivable that HGNA shear zones will be active over long periods of time or that the same boundary conditions will re-occur at the same or different levels, at different times. This could result in extreme diachronism of structures that would be grouped, on the basis of style and overprinting, into common generations.

Compositional layering, being in general a transposition foliation in high-strain zones, cannot be assumed to represent the flow extensional eigenvector of the zone. Making such an assumption in determining the kinematic vorticity number of natural shear zones may underestimate the bulk kinematic vorticity number of the zone and lead to incorrect conclusions regarding the significance of the pure-shear component in the zone.

Palinspastic reconstructions based on the assumption that the thrust-like geometry is a product of thrusting, are likely, at best, to be highly misleading if applied to mid- to lower-crustal rocks.

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