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Geometric and kinematic evolution of asymmetric ductile shear zones in thrust sheets, southern Adelaide Fold–Thrust Belt, South Australia

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Abstract

The Southern Adelaide Fold–Thrust Belt, South Australia contains unusually asymmetric shallow to moderately inclined low-grade phyllonitic shear zones, uniformly floored by basal imbricate thrusts. The fabrics and minor structures that reveal the geometric and kinematic evolution of these shear zones are described, analysed and presented in detail. Incipient fabrics and structures are developed in upper transitional zones and then develop progressively more intensely downwards into the shear zones. They ultimately intensify along the lower boundary thrusts that make up the base of all of the shear zones. Microstructural analysis of multiple cleavages and grain-scale geometries from within the shear zones demonstrate that deformation is spatially and temporally concentrated along the lower boundary thrusts. The shear zones initiated by strain-softening at the base of a number of stacked thrust sheets during Mid- to Late-Cambrian orogenic shortening of the area when the fold–thrust belt was generated. Kinematic indicators confirm consistent NW-directed transport. The shear zones progressively developed upwards by further superposition of shear and flattening strain during which time deformation was localised within the base of the thrust sheets to develop distinctively asymmetric shear zones. © 2000 Elsevier Science Ltd. All rights reserved.

1. Introduction

Thrust faults and shear zones contain some of the key information to understand the evolution of fold–thrust belts (Mitra and Fisher, 1992; McClay, 1992). Detailed structural mapping, in conjunction with studies of the regional development of deformation fabrics and structures associated with thrusts/shear zones, have contributed much to understanding the development of classical fold–thrust belts such as the Appalachian Valley and Ridge province (Mitra, 1988; Mitra and Wojtal, 1988) and Canadian Rocky Mountains (Price, 1981; Fermor and Price, 1987). Thrusts and shear zones from the external parts of orogenic belts (especially fold–thrust belts) are generally considered as discrete and narrow faults because they initiate as brittle fractures (Hubbert and Rubey, 1959; Hsü, 1969; Muller and Hsü, 1980). In the more internal portions of orogenic belts, however, large weakly to moderately deformed sheets of rock commonly move by concentrating shear strain in a narrow zone at their base, which leads to the formation of mylonites and arrays of associated minor structures (Harris and Milici, 1977; Schmid, 1983; Gilotti and Kumpulanien, 1986; Wojtal, 1986; Wojtal and Mitra, 1988).

In the southern Adelaide Fold–Thrust Belt (SAF–TB), basin inversion and contractional deformation are expressed by regional low- to moderate-angle thrusting (Flottmann and James, 1997), commonly involving basement slices. These structures comprise a major stacked array of basal thrusts and associated

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overlying asymmetric shear zones. Here we review the structural and strain characteristics of ductile shear zones that lead to the characteristic deformational style in such thrust sheets. We provide a detailed examination of the geometry, kinematics, microstructure, deformation mechanisms, and finite strain of the thrusts and shear zones in the belt. Based on these analyses, we demonstrate and discuss a model for the formation of these thrusts and shear zones, which may serve as a case study for less well exposed thrust belts in other parts of the world.

2. Structural interpretations of the SAF–TB

The Adelaide Fold–Thrust Belt is a major foreland fold–thrust belt formed by the Cambro-Ordovician Delamerian Orogeny of South Australia (Flottmann et al., 1994; Marshak and Flottmann, 1996; Flottmann and James, 1997). The southern section of the Adelaide Fold–Thrust Belt in south Australia (SAF–TB, see Fig. 1) was formerly regarded as a classical fold belt, governed by a single large-scale folding event with involvement of basement in the cores of major anticlines (Sprigg, 1946; Offler and Fleming, 1968; Anderson, 1975; Mancktelow, 1981, 1990). Thrusts in the belt were considered as minor structures which only occurred on the steep to overturned west-facing limbs of the major asymmetric F_1 folds, which formed during the D_1 regional folding (Offler and Fleming, 1968; Mancktelow, 1990). The early lack of recognition of thrusting as an important process was largely because of the deficiency of recorded evidence for emergent faults and large-scale displacements (thrusts or shear zones) at the surface.

Structural mapping on the Fleurieu Peninsula and Kangaroo Island has shown the existence of many thrusts and shear zones in the SA–FTB (Flottmann et al., 1994, 1995; Flottmann and James, 1997). However, there was still uncertainty as to the location and continuation of these thrusts and shear zones into the area of the Adelaide Hills and Mount Lofty Ranges (Fig. 1). Subsequently, detailed structural mapping across the Adelaide Hills has revealed the location of many more thrusts and shear zones in the northern parts of the SA–FTB. It has also allowed recognition of the con-

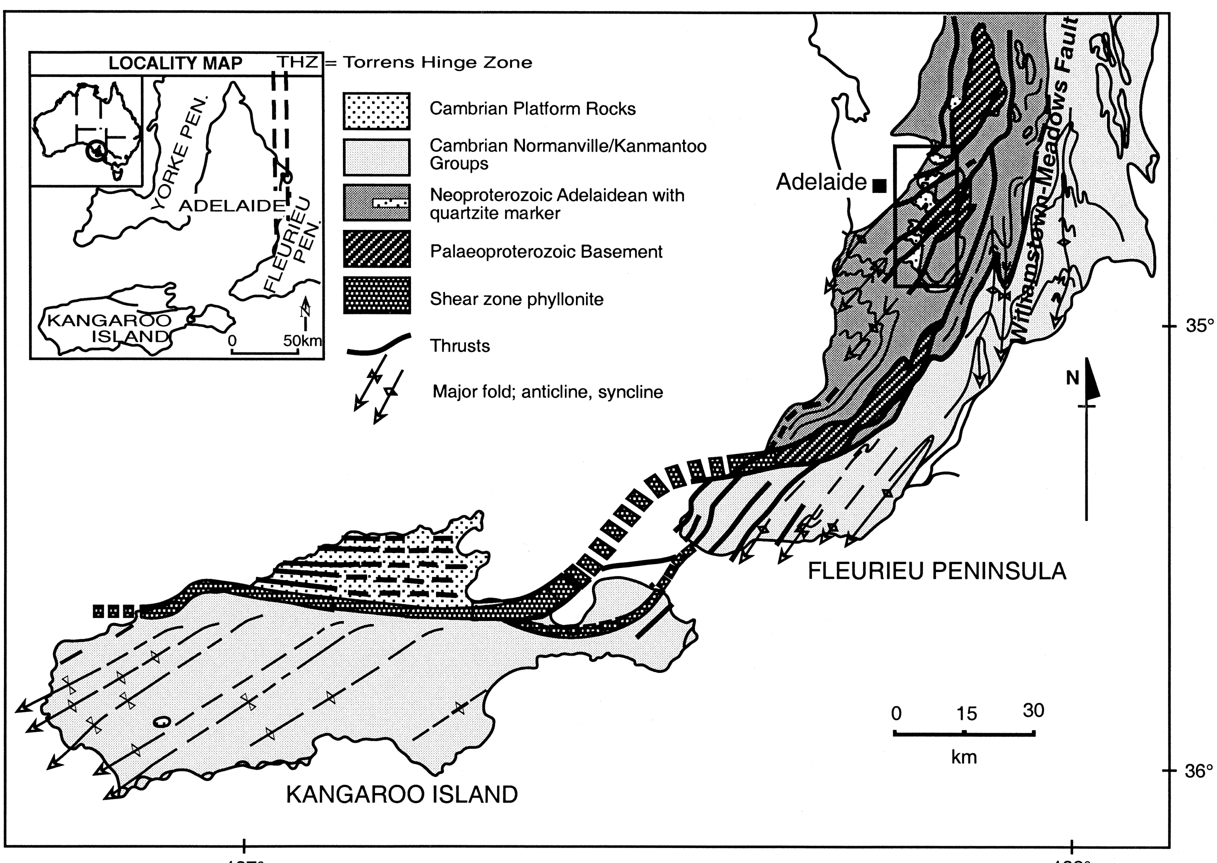

Fig. 1. Simplified geological map of the southern Adelaide Fold–Thrust Belt; modified after Flottmann et al. (1994) and Flottmann and James (1997). The location of the southern Mt. Lofty Ranges/Adelaide Hills area directly east of Adelaide is shown within the box (see Fig. 2 for more detail).

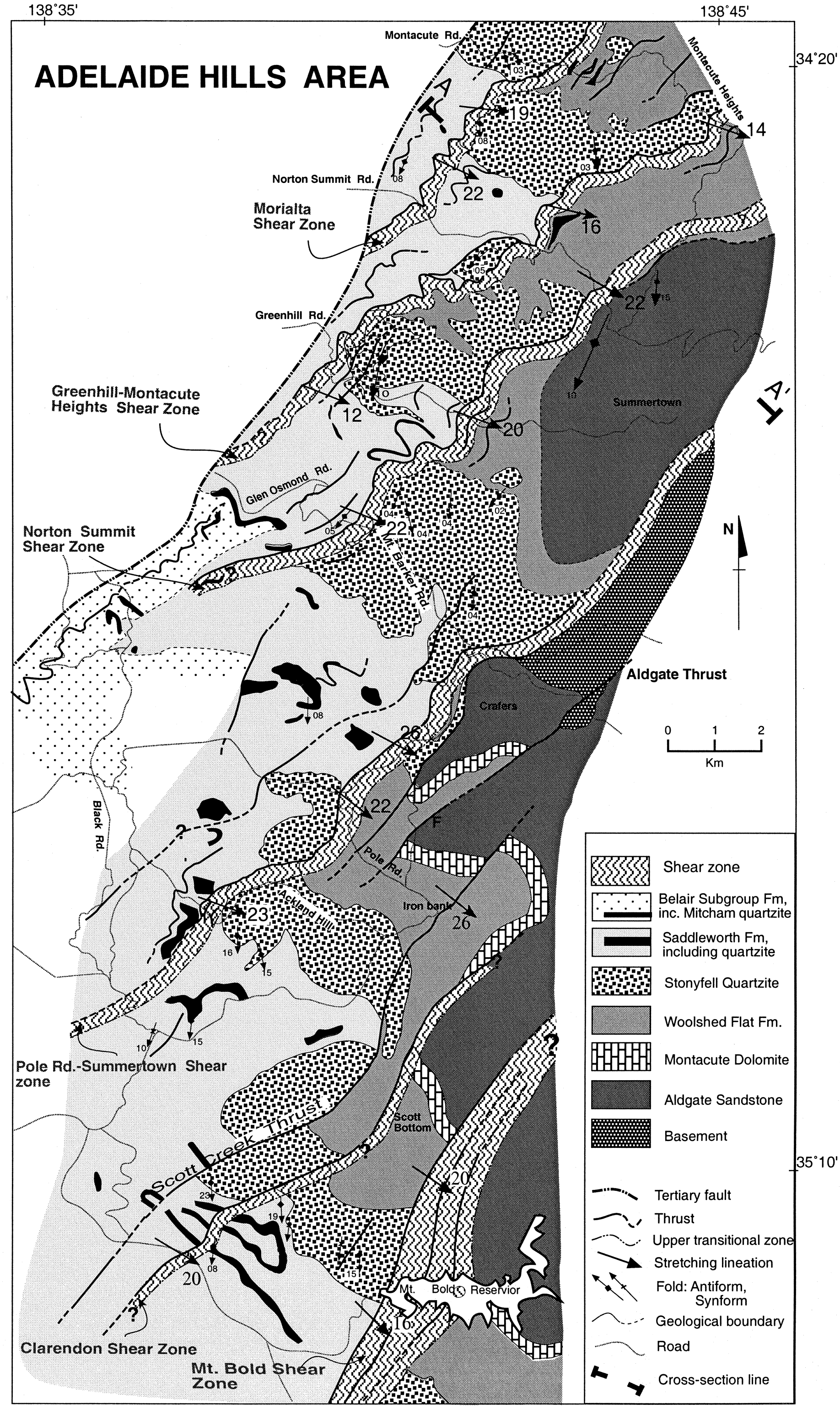

Fig. 2. Geological map of the Adelaide Hills area showing the location of ductile thrusts and shear zones, and the cross-sections of Fig. 4. Map is compiled from data of geological survey mapping (Forbes, 1979, 1980, 1983) and from mapping by Yassaghi (1998), Gessner (1996), Lill (1996), Markwitz (1998) and Weil (1998).

tinuation of some thrusts and shear zones which penetrate higher stratigraphic levels from the Fleurieu Peninsula northwards into the Mount Lofty Ranges (Yassaghi, 1998).

The Adelaide Hills area is a part of the southern Mt. Lofty Ranges and is situated in the northern part of the SAF–TB. In the area a number of discrete Palaeoproterozoic basement inliers are overlain by a sequence of Neoproterozoic Adelaidean mainly clastic low grade metasedimentary rocks of the Burra Group (Preiss, 1987). The Burra Group in this area, based on previously published geological maps (Forbes, 1979, 1980, 1983), comprises mainly multiple units of competent sandstone and quartzite of the Aldgate Sandstone and Stonyfell Quartzite interbedded with incompetent pelitic rocks of the Woolshed Flat and Saddleworth Formations (Fig. 2). The Burra Group has a maximum thickness of about 3000 m, much of which is the Aldgate Sandstone that directly overlies the basement inliers. Metamorphic conditions across the area are uniform (biotite grade/greenschist facies) and show that progressive deformation took place at or near the brittle–ductile transition. Both fluid inclusion studies in quartz veins and microprobe analyses of micas in

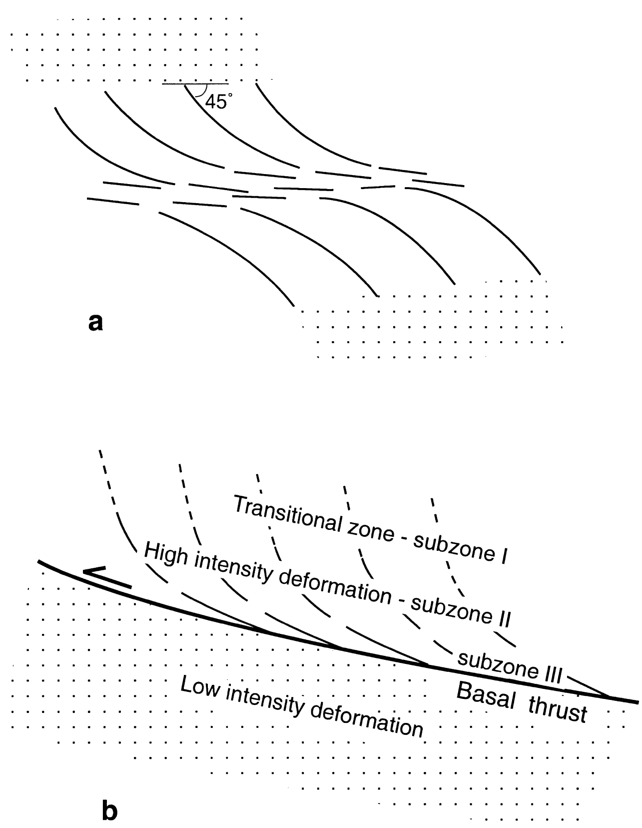

Fig. 3. Cartoons showing (a) a typical ductile shear zone with a maximum displacement gradient in the shear zone centre; (b) a typical asymmetric shear zone from the Adelaide Hills area with a distinct basal thrust.

pelites show average peak temperatures of 350–400°C at 3.5–4.0 kbar (Yassaghi, 1998).

Detailed structural mapping has revealed six major stacked parallel thrust sheets generally trending NE–SW with shallow SE dips (Fig. 2). It is the documentation of these structures that are presented and discussed in detail here.

3. General characteristics and distribution of the thrusts and shear zones

The most notable deformation features of the Adelaide Hills area are the well-developed fabrics and structures in narrow subparallel deformation zones, separated by wider areas with less concentration and intensity of structures. The geometry of the shear zones are unusual in that they are not like typical ductile shear zones in crystalline rocks (Ramsay, 1980), which are symmetric with sygmoidal fabrics and with displacement gradients decreasing symmetrically towards both margins from a maximum in the centre (Fig. 3a). Shear zones in the Adelaide Hills area are rather asymmetric with distinct basal discontinuities or thrusts (Fig. 3b). However, the presence of asymmetric structures and shear bands, and the gradual curvature of the foliations, implies that these high deformation zones or shear zones are governed by noncoaxial deformation regimes.

The shear zones strike generally NE–SW (040–050°) and dip moderately SE (25–40°) (Fig. 2). Their mapped strike-lengths range from 8 to 20 km, although the poor outcrop makes it difficult to follow them. The map widths of the shear zones vary from 250 m (Clarendon Shear Zone) to more than 1 km (Mt. Bold Shear Zone), depending mainly upon the competence of the rock units and localisation of deformation. The upper boundaries to the shear zones are transitional and it is difficult to constrain a distinct boundary, whilst the lower boundaries to the shear zones are generally sharp and bounded by discrete basal thrusts (Fig. 3). As the lower boundary thrusts reach the surface they are therefore emergent lower boundary thrusts (Boyer and Elliot, 1982).

The shear zones display several features analogous to those recognised in conventional thrust zones (Boyer and Elliot, 1982; Evans, 1989) with older rocks generally emplaced over younger rocks. The stretching lineations in the shear zones generally trend southeast, which together with overthrust kinematic indicators imply northwestward displacement across the shear zones (Fig. 4). The orientation of this lineation varies between the competent and incompetent units; and it trends more easterly in competent units.

The shear zones represent the loci of emergence of a major crustal discontinuity or detachment possibly extending to the middle crust (Fig. 4), and are responsible for the exhumation of the basement inliers and for their involvement in the shearing throughout the SA–FTB. Similar basement slices are also thought to underlie the unexposed cores of major ramp anticlines formed by large areas of exposed Aldgate Sandstone, mainly seen in the eastern part of the area around Summertown, Crafers, and east of Mt. Bold (Fig. 2).

All of the shear zones anastomose and bifurcate to produce separate and distinct arrays of minor thrust-like shear zones that form foreland-verging imbricate fans verging to the NW. These narrow zones of ductile deformation contain the same array of fabrics and deformation structures with the same geometry and kinematics as developed in the major shear zones. They are formed either within the shear zones (e.g. Mt. Bold shear zone, Fig. 2) or in the footwalls and hanging walls to the shear zones (e.g. Greenhill and Norton Summit shear zones, Fig. 2). The lengths, widths, and intensities of deformation of these thrusts are variable but are generally considerably less than those of the shear zones. The thrusts with the longest trace lengths (Scott Creek Thrust and Aldgate Thrust, Fig. 2) have wider zones of deformation with greater development of fabrics and structures than the smaller ones.

Folds, mostly anticlines with overturned western limbs, are commonly observed in the hanging walls of the lower boundary thrusts to the shear zones and also occur along the associated imbricate thrusts. Most folds plunge shallowly (usually less than 10°) south-southwest and they are generally best developed in more competent quartzite units. The geometry of these folds closely resemble the geometry of fault-propagation folds and in almost all cases the shear zones and associated thrusts were developed on the steep to overturned forelimb of these folds and produce a structure similar to the geometry of synclinal break-through fault propagation folds (Suppe and Medwedeff, 1990). It is most likely that the minor folds found within low intensity deformation areas between the shear zones developed at the tips of blind thrusts at depth.

4. Geometry, kinematics and strain in the shear zones

Having demonstrated the presence of asymmetric shear zones in this area, it is important to resolve the physical processes inherent in their development. In this section we present the detailed physical characteristics of fabrics and structures developed across the thrusts and shear zones and analyse them descriptively and statistically in order to document their unusual character and interrelationships, and to define their geometry and kinematics. The main geometrical elements recognised to occur specifically within the shear zones are bedding (S_0), arrays of composite and overprinted foliations (S_{1a}, S_{1b}, and S_{1c}), minor folds, extensional crenulation cleavages (ECC-fabrics of Passchier, 1991), deformed quartz veins, and stretching lineations.

4.1. Planar fabrics and cleavages

Bedding (S_0) and to a lesser extent a slaty cleavage of variable intensity (recognised as the regional cleavage, S_{1a}) are the dominant fabric elements outside of the shear zones. The S_{1a} cleavage is the earliest recognisable tectonic structure, and is a weak to moderate continuous cleavage (Powell, 1979). It is most strongly developed within incompetent pelites, carbonates and occasionally in psammitic rocks. The S_{1a} cleavage

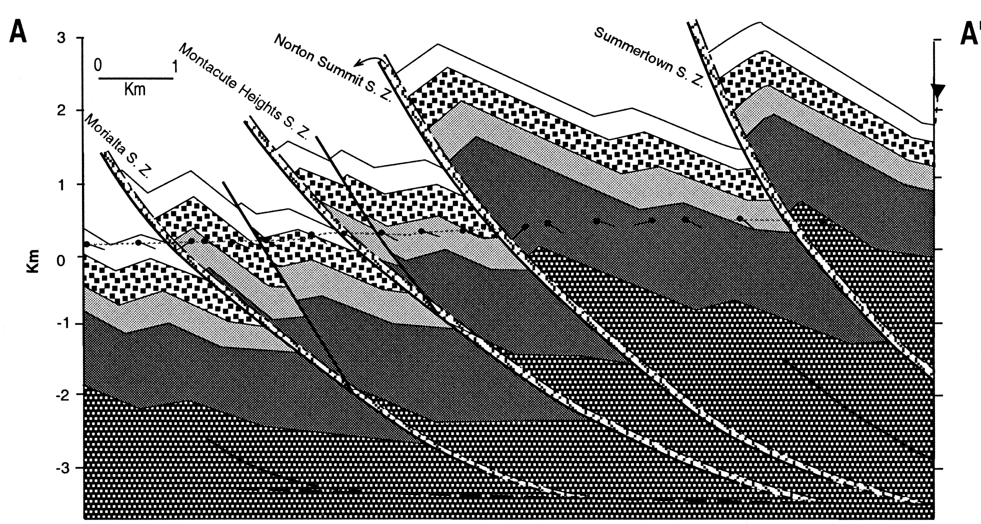

Fig. 4. Cross-section across the Adelaide Hills area. For the location of the section line (AA′) see Fig. 2.

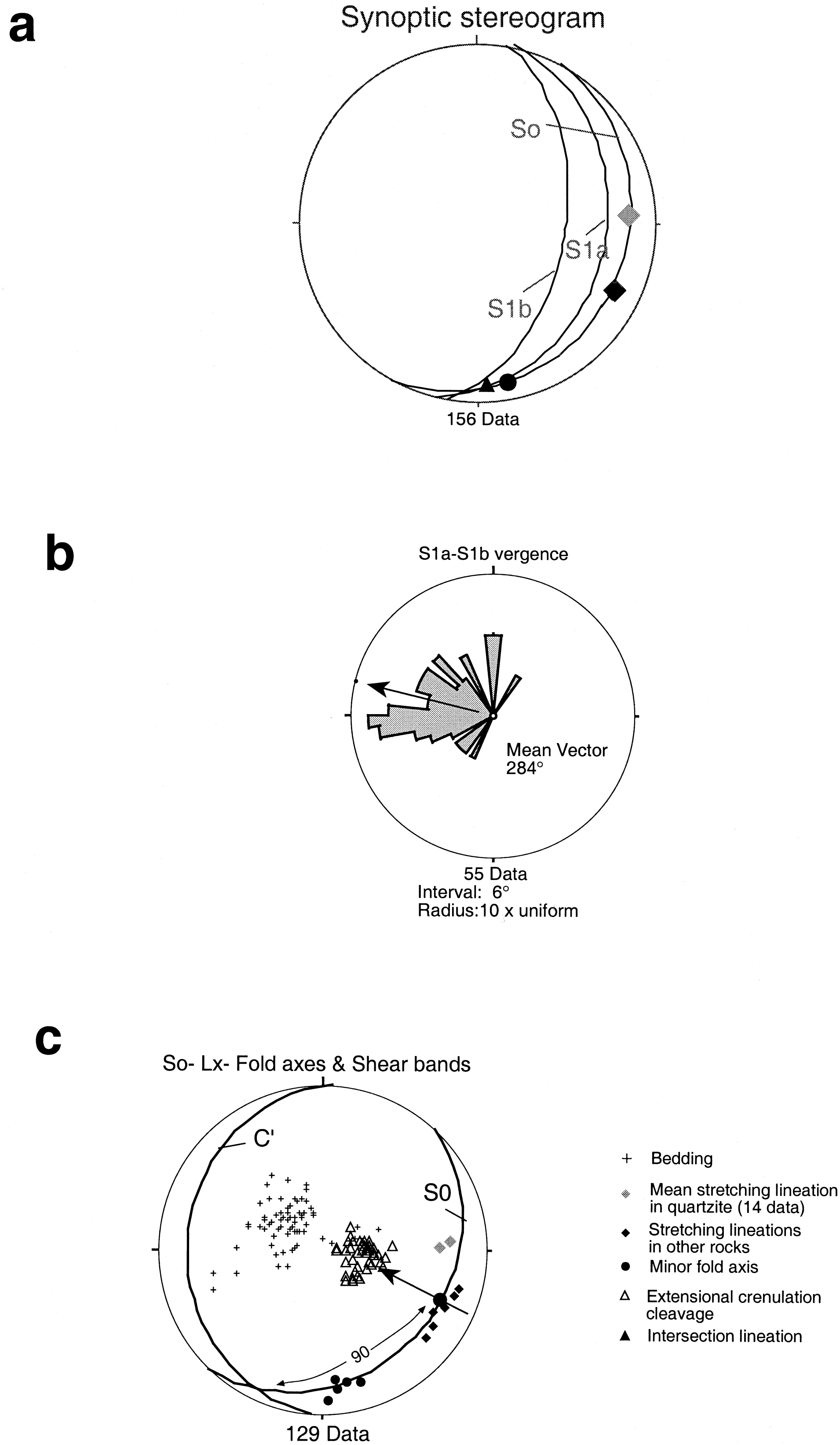

Fig. 5. Stereograms and rose diagram showing transport direction analysis of the shear zones. (a) Synoptic stereogram from the area adjacent to Greenhill road, showing the consistent relationship between S_0, S_{1a}, and S_{1b}. (b) Vergence direction analysis calculated as azimuth direction perpendicular to intersection of overprinting foliations. (c) Synoptic stereogram of selected data from the Adelaide Hills area. The arrow shows constructed movement direction of the shear zone based on the intersection of the great circles of mean orientations of the shear bands (C') and bedding planes (S_0), which represent orientation of shear planes. For more explanation see text.

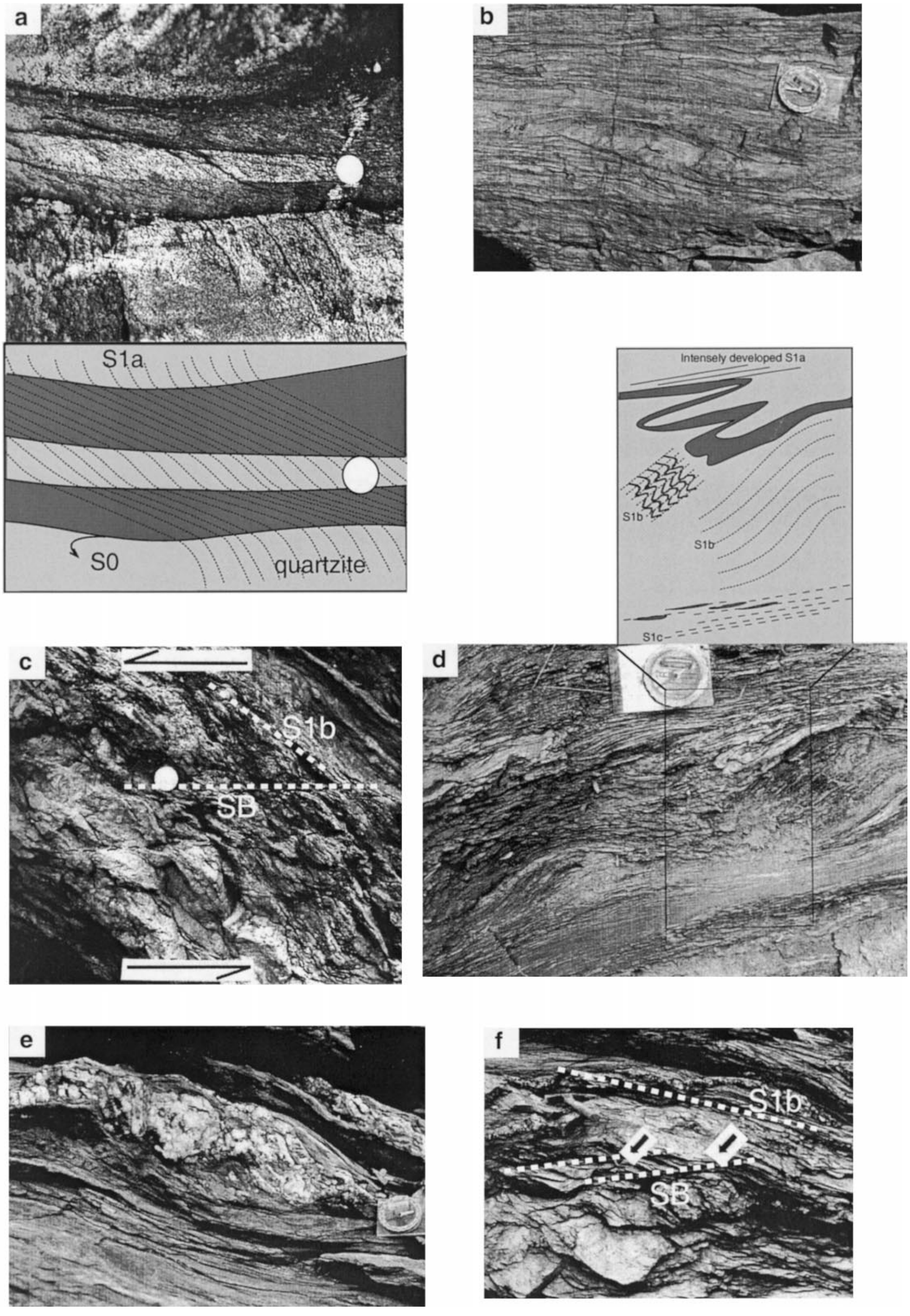

Fig. 6. Photographs and sketches of minor structures. (a) Interlayered psammopelites and psammites in thick quartzite of the Stonyfell Quartzite in the Morialta area. The cleavage dies out within the thick quartzites on both sides. In the sketch also note the refraction of S_{1a} from psammopelites into quartzitic psammites; the coin diameter is 1.5 cm. (b) Transposition of bedding in psammites from the Woolshed Flat Shale within the Mt. Bold shear zone. The asymmetric geometry of the boudin like structure is consistent with the northwest sense of movement (to the left). (c) Sheared Stonyfell Quartzite about 50 m above the lower boundary thrust to the Morialta shear. Note the development of shear bands or ECC-fabrics (dashed lines) which cause offset/extension of the lozenge-shape domains to the left (northwest); the coin diameter is 1.5 cm. (d) Development of cleavages from S_{1a} to S_{1c} in phyllonites of the Woolshed Flat Shale from within the Clarendon shear zone (looking southwest). (e) Asymmetrically boudinaged quartz veins from within the Mt. Bold shear zone where the sense of movement can be shown towards northwest (to the left). (f) Asymmetrical extension shear bands (arrows) in phyllonites of the Woolshed Flat Shale from within the Mt. Bold shear zone.

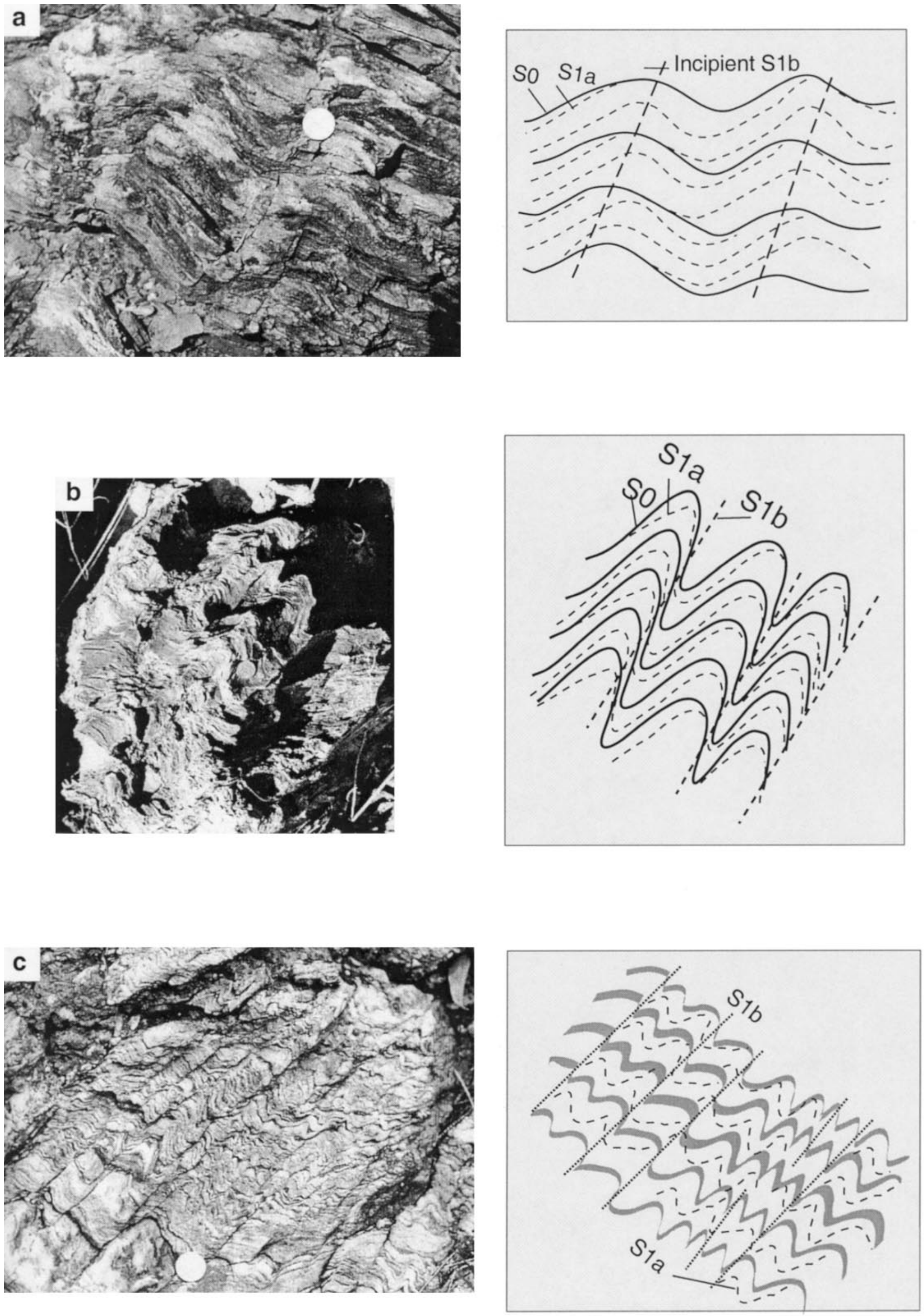

Fig. 7. Progressive development of overprinting cleavages from the upper transitional zone toward the lower boundary thrust in the Morialta shear zone. Because of weak development of the S_{1a} cleavage in pelites, it is not easy to see S_{1a} in these photographs. (a) Development of weak (open) crenulation folds at the upper transitional zone where earlier regional cleavage (S_{1a}) is deformed to develop a new crenulation cleavage (S_{1b}). (b) Crenulation folds and cleavages within the shear zone; note the initiation of dissolution surfaces on the short steeper forelimbs of the crenulation folds. (c) Typical crenulation cleavages (S_{1b}) in the phyllonites about 20 m from the lower boundary thrust where the dissolution surfaces are well developed (see also the sketch). The diameters of the coins are 1.5 cm.

forms an axial plane fabric to the regional shallowly inclined folds, strikes N-NE to S-SW and dips SE, and typically lies at low angle to, and is generally more steeply dipping than, the bedding (Fig. 5a). The overturning direction of the cleavage is consistently to the W-NW, irrespective of the dip of the bedding (Fig. 5b). In competent quartzite units, S_{1a} is recognised by an alignment of phyllosilicates in thin incompetent psammitic and pelitic units interbedded within thick to massive quartzite. In these interbedded units the cleavage is not pervasive and refracts from pelites to psammites (Fig. 6a).

Evidence of shearing is seen approaching the upper transitional zone with the formation of a weak disjunctive cleavage in the quartzite units. Cleavage domains are spaced at intervals of about 2–5 cm, and occupy less than 10% of the rock. Within the incompetent pelites, the onset of shearing is defined by the development of open, low amplitude (5–10 cm) crenulation folds with asymmetric shapes and linear hinges spaced at intervals of about 10–20 cm (Fig. 7a).

Within the shear zones, the regional cleavage is intensified and more strongly developed in almost all lithological units, but is most significantly enhanced in the incompetent units. The cleavage is deformed to form crenulation cleavages in pelites and phyllites and/or transposition cleavages in pelites and psammites (Fig. 6b). These superposed cleavages are termed S_{1b} to differentiate them from the regional cleavage. In the quartzite units, S_{1b} is a weak cleavage defined by the alignment of slightly elongate quartz grains to form a spaced cleavage. The spacing between the cleavage surfaces is generally about 2 cm but not very regular. At about 50 m from the lower boundary thrusts this spaced cleavage is more strongly developed and pervasively foliates the thick layered quartzite (Fig. 6c). Ultimately this cleavage is intensely developed with about 5-mm spacing.

Similarly, in the incompetent units, the open zonal crenulation folds within the transitional zones (Fig. 7a) are gradually more intensely developed into the shear zones to form crenulation folds (Fig. 7b). These crenulation folds, which are defined by microfolding or crenulation of the pre-existing layering (S_0) and slaty cleavage (S_{1a}) are asymmetric and have linear hinges spaced at intervals of 5–10 cm. Bedding becomes progressively transposed with thinning of limbs and formation of relict rootless fold hinges. The microfold limbs of the crenulations are often sheared out and show dark residual seams and truncations as evidence of pressure solution and redistribution of minerals. These dissolution surfaces, which almost always form on the steep to overturned short limbs of the asymmetrical microfolds, and sometimes cut across earlier axial planes, define a weak cleavage similar to the crenulation cleavage S_{1b}. Crenulation folds are more frequently developed in finely laminated pelites and to a lesser extent in psammites. Closer to the lower boundary thrust (20 m from the thrust), the crenulation folds are further intensified to form strong S_{1b} crenulation cleavages (Fig. 7c), with orientations close to those of S_{1a}, but consistently steeper (Fig. 5a). As is shown in Fig. 7(c), S_{1b} is a dissolution cleavage (Gray, 1979) and in some outcrops, where the rate of dissolution was higher, S_{1a} has simply rotated to form S_{1b} without the intervening crenulation phases (see also Tobisch and Patterson, 1988).

At the lower boundary thrusts to the shear zones, the crenulation cleavages (S_{1b}) are completely transposed to form a new continuous cleavage S_{1c} (Fig. 6d), whose orientation is sub-parallel to S_{1a} (see sketch map of Fig. 6d). This cleavage is often difficult to distinguish from the initial slaty cleavage (S_{1a}) in outcrop or even in thin sections. Progressive development of cleavages from S_{1a} to S_{1b} and finally to S_{1c} within the shear zones appears to be gradational and uniform through all the shear zones, although it is not always possible to continuously follow this development due to lack of exposure.

4.2. Stretching lineations

A stretching lineation is a consistently developed fabric element in the shear zones and associated basal thrusts. The lineation is defined by minerals (quartz and chlorite) and mineral aggregate preferred alignment, pressure fringes around clasts and stiff crystals and other grain elongation. The lineation in the quartzite units plunges shallowly east, while in the pelite and phyllite units it plunges closer to southeast (Fig. 5a). It has been recognised that where different layers have different viscosities, i.e. planar anisotropy, the orientation of the stretching lineation within incompetent layers generally lies closer to the true direction of tectonic transport (Ridley, 1986). This is demonstrated in the current area by the independent interpretations of vergence directions and shear plane analyses. These analyses show that the transport direction of the Adelaide Hills shear zones is towards the W-NW (with an average of 287°) (Fig. 5b and c). This movement direction is coincident with the movement direction implied from the orientation of the stretching lineations in the incompetent units, i.e. toward the northwest. These interpretations are also in accordance with the northwest sense of movement obtained from asymmetric minor folds (Fig. 8a), asymmetric tails of boudinaged quartz veins (Fig. 6e), and extensional crenulation cleavages or shear bands (Fig. 8g). The trend of the stretching lineations in the competent quartzite units, which is more easterly, reflects active bulk asymmetric anticlockwise rotation of these rigid units towards the movement direction during deformation

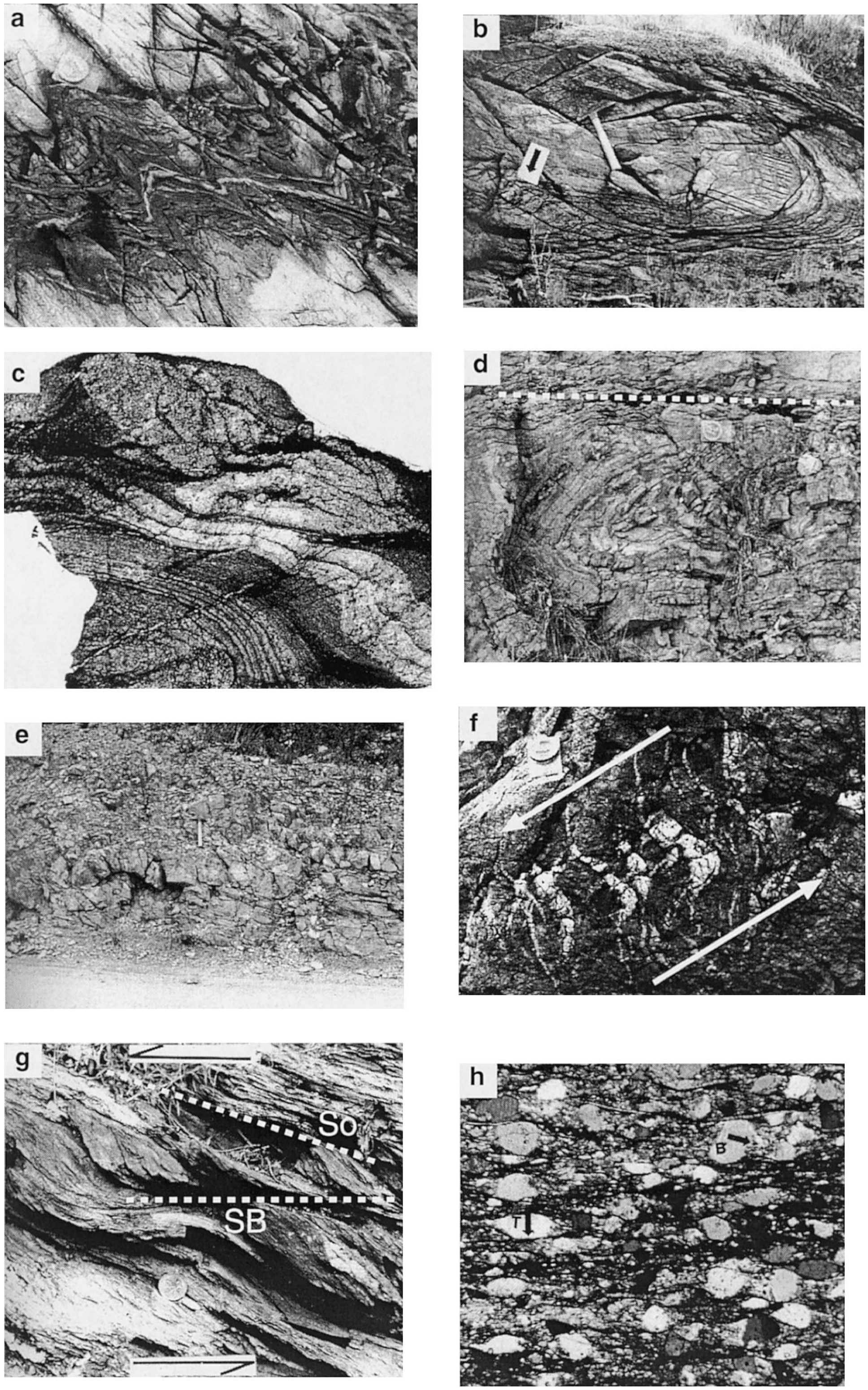

Fig. 8. Photographs of minor folds and fabrics from within the shear zones. (a) Minor asymmetric Type II fold in the Woolshed Flat Shale from within the Mt. Bold shear zone. Note the sub-parallelism of the axial plane with S_{1b}. (b) Isoclinal recumbent sheath-like folds (Type I) in the Woolshed Flat shale from within the Mt. Bold shear zone, where characteristically its upper limb is truncated by thrusting. Also note superposition of minor asymmetric Type II folds on the lower limb (arrow). (c) Photomicrograph (*YZ* section) of Type I fold geometry showing sheath-like nature of these folds. Notice eye fold closure in the upper part of photomicrograph. (d) Thrust truncation of the upper limbs of Type I fold within the Mt. Bold shear zone (looking southeast). (e) Out of syncline thrust geometry from within the Mt. Bold shear zone, see text for more detail (looking southwest). (f) Development of en-échelon quartz veins in the Stonyfell Quartzite from the Morialta shear zone. Note that the veins were deformed progressively in a semi-ductile manner in which more than one generation of the veins can be seen. (g) Extensional crenulation cleavage (ECC-fabrics) in the phyllonite from within the Clarendon shear zone; the coin diameter is 1.5 cm. Note the angle of $\alpha = 30^\circ$ between bedding or the *C* planes (dashed lines) and the shear bands. (h) Well-developed rough cleavage in psammites of the Woolshed Flat Shale from within the Mt. Bold shear zone. Here, the cleavage domains are more closely spaced (100–300 μm), more continuous, and thicker seams. Quartz grain boundaries are either planar truncated 'T'. Overgrowing mica beards 'B' have also developed on larger and more spherical grains; crossed polars; width of view = 8 mm.

(Sanderson, 1973). The trend of the stretching lineations within the incompetent units, also changes slightly (from 109° to 130° average) from the northwestern shear zones (the Morialta shear zone, Fig. 2) to the southeastern Mt. Bold shear zone (see also Fig. 5a—Greenhill Rd.–Pole Rd. zones), reflecting a gross overall variation in movement, most likely caused by buttressing closer to the foreland against the north–south-trending belt margin of the Torrens Hinge Zone.

4.3. Minor folds

Minor folds are ubiquitous in the shear zones and are generally disharmonic structures forming open-to-tight buckle-style folds plunging shallowly towards the S-SE (Fig. 5a). They mainly occur where the rocks within the shear zones are anisotropic, due to interbedding of psammites and pelites. Their geometry, however, is not consistent across the shear zones and two types of minor fold geometry have been distinguished. The first type (Type I) includes moderately to strongly curvilinear hinges, close-to-isoclinal near recumbent fold styles with fold axes sub-parallel or at low angles to the local stretching lineation (Fig. 8b). Folds of this type are best developed within the Mt. Bold shear zone where they plunge shallowly southeast and show a sheath-like geometry (Fig. 8c). The upper limbs of these folds are consistently truncated and cut-off by small-scale minor imbricate thrusts (Fig. 8d), which are possibly related to the occurrence of out-of-sequence thrusting and shearing.

The second fold type (Type II) includes gentle-to-tight asymmetric folds with long gently dipping southeast limbs and short steeply overturned northwest limbs (Fig. 8a). Folds of this type plunge shallowly S-SE (Fig. 5a) and their axial planes consistently dip moderately southeast. These folds display little or no detectable hinge line curvature and the fold axes are almost always oblique to the local stretching lineation (Fig. 5a). The length ratio of long limbs to short limbs of these asymmetric folds varies across each shear zone. Near the lower boundary thrusts, the short limbs of the folds are almost always sheared-off and/or apparently displaced probably due to effects of dissolution processes. At the lower boundary thrusts, small scale imbricate thrusts propagate through these sheared-off limbs (Fig. 8e), and carry the antiforms some distance from the syncline in the footwall, creating small-scale out-of-syncline thrust geometries (Coward, 1988). The Type II folds are superimposed on the earlier isoclinal-recumbent Type I folds (see the arrow on the lower left side of Fig. 8b), which together with their close spatial relationship to the shear zones, suggests progressive superimposed folding during the gradual development of the shear zones rather than during discrete overprinting deformation phases.

4.4. Quartz veins

Lens-shaped quartz veins occur in the shear zones with major and intermediate axes almost always subparallel to S_0 and/or S_{1a}. They are mostly seen in the phyllites and to a lesser extent in the quartzite within the shear zones and along the associated imbricate thrusts (Fig. 6f). The majority of quartz veins are spatially confined to the shear zones indicating a clear genetic relationship between the two. They are often more intensely deformed to form boudinaged veins. These boudinaged veins are almost always asymmetric with asymmetric tails (Fig. 6e), indicating a sense of shear towards the northwest. The sheared quartzite in the Morialta Shear Zone contains numerous semi-ductile en-échelon quartz vein arrays (Fig. 8f), whose orientations are also consistent with a top-to-the-northwest shear zone movement sense.

4.5. Extensional crenulation cleavages (shear bands)

Composite planar fabrics (shear bands) in the form of the extensional crenulation cleavages (ECC-fabrics of Passchier, 1991) also occur within the shear zones (Fig. 8g). These fabrics strike mainly N-NE and dip shallowly ($<20°$) W-NW (Fig. 5a). Unlike the minor folds and crenulation cleavages, which are more intensely developed towards the lower boundary thrusts to the shear zones, the shear bands are best developed within the shear zones but closer to the upper transitional zones. The concentration of these fabrics decreases substantially towards the lower boundary thrusts. They occur mostly in phyllites on centimetre scales (up to 20 cm). They also occur in highly sheared quartzite where the spaced cleavages in quartzites are anastomosing and divide the rocks into lozenge-shaped domains. These domains are separated along shear planes and demonstrate the geometry of the shear bands (Fig. 6c). As is shown in Fig. 8(g), the shear bands in the Clarendon shear zone make an angle of 30° to the shear planes, which are considered to be subparallel to the bedding. This angle varies from 25° in the Mt. Bold shear zone to 40° in the Pole Road shear zone. This suggests that the intensity of shearing in the Mt. Bold shear zone appears to be greater than those of the Pole Road and Morialta shear zones. As is also shown in Fig. 8(g), the shear bands that dip shallowly northwest cause extension of the S_{1a} slaty cleavage, which dips to the southeast. From this dip direction, the sense of shearing can be defined (Simpson and Schmid, 1983; Lister and Snoke, 1984; Hanmer and Passchier, 1991) and is consistently toward W-NW.

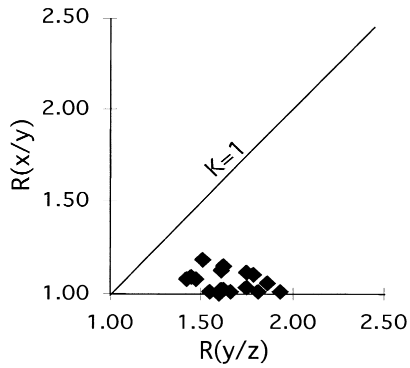

Fig. 9. Flinn diagram for finite strain geometry of the Stonyfell Quartzite samples across the shear zones in the Adelaide Hills area.

4.6. Strain analysis

Finite strain analyses on detrital quartz grains from quartzite units was carried out using the R_f/ϕ technique (Fig. 9). All samples show low minimum strain values, with a general trend to apparent flattening strains. Samples from upper transitional zones show low *XZ* strain magnitudes (1.6–1.8), while those from close to the lower boundary thrusts are higher (~2.0), but all values reflect the influence of crystallisation or solution processes rather than ductile distortion (see Section 5).

5. Microstructures of the shear zones

Detailed microstructural studies including fabric development and integrated grain-scale microstructural changes were carried out to help constrain the spatial and geometric development of the shear zones.

5.1. Microstructural development of cleavages in the shear zones

The geometric development of the shear zones and the deformation mechanisms revealed by cleavage development was investigated from thin sections collected on detailed transects across the shear zones and lower boundary thrusts. Detailed microstructural studies of rough cleavage (Gray, 1978) developed in psammites from across the shear zones reveal three main microstructural changes. These are: (1) the grain size of detrital quartz grains decreases from 200–600 μm to 50–500 μm; (2) the proportion of needle-like and syndeformational white mica, which produces the rough cleavage, increases (compare Fig. 10a with c). The volume percentage of those grains which produce the rough cleavage increases from less than 1% of the rock volume (Fig. 10a) in samples from outside the shear zones, to more than 30% of the rock volume in samples from the lower boundary thrusts to the shear zones (Fig. 10c); and there is (3) an increase in the amount of intracrystalline deformation features in detrital quartz grains. Quartz grains in samples from the upper transitional zones show only weak undulatory extinction while samples from the lower boundary thrusts exhibit evidence of sweeping undulatory extinction, deformation bands, and in grains with higher dislocation density, even subgrain formation and small recrystallised grains (Fig. 10b).

5.2. Microstructures of quartzite in the shear zones

The detailed study of microstructures in Stonyfell Quartzite from all shear zones shows that from outside through the upper transitional zone and toward the lower boundary thrust, the crystal-scale deformation increases. Quartz grains have undergone crystal-plastic deformation, while feldspar grains deform by intracrystalline fracturing. The amount of intracrystalline fracturing of feldspar grains also increases within the shear zones and towards the lower boundary thrusts and between the shear zones from the Morialta to the Mt. Bold shear zones.

Outside of, and in the upper transitional parts of, the shear zones, quartz grains are equant polygonal to slightly elongate and their grain boundaries are bulged. Oriented growth of small white micas on the margins of quartz grains is also seen, especially in the Morialta shear zone (Fig. 10d). The grains also exhibit intracrystalline deformation features including patchy to slightly undulatory extinction. The types of quartz microstructures from within these parts of the shear zones are coincident with microstructural development of experimentally deformed quartz at the first stage of shearing (Simpson, 1994) or in regime 1 of Hirth and Tullis (1992).

Within the shear zones, quartz grains exhibit sweeping undulatory extinction, deformation bands, subgrain formation and small recrystallised grains on the margin of original grains with increased dislocation density (Fig. 10e). This type of microstructure is similar to the microstructures developed in regime 2 of the experimentally deformed quartzites of Hirth and Tullis (1992), in which a subgrain rotation microstructure is suggested to be responsible for recovery at moderate temperatures or slow strain rates (Urai et al., 1986; Hirth and Tullis, 1992; Simpson, 1994).

At the lower boundary thrusts to the shear zones, the amount of recrystallised small grains on the margins of original grains is enhanced (Fig. 10f) and leads to development of core and mantle structures (White, 1976). These microstructures are almost identical to microstructures developed in regime 3 of Hirth and Tullis (1992) in experimentally deformed quartzites in which dislocation climb is sufficiently high to control recovery at high temperature or lower strain rate.

5.3. Quantitative analysis of microstructural quartz grain geometry

A study of the variation of grain size, grain shape (aspect ratio) and shape preferred orientation (ϕ) was carried out to demonstrate and tentatively quantify development of microstructures from within the Stonyfell Quartzite units in the shear zones (Figs. 11 and 12). Measurement of the parameters was carried out on photomicrographs, which were taken from the same oriented (XZ) thin sections as were used for

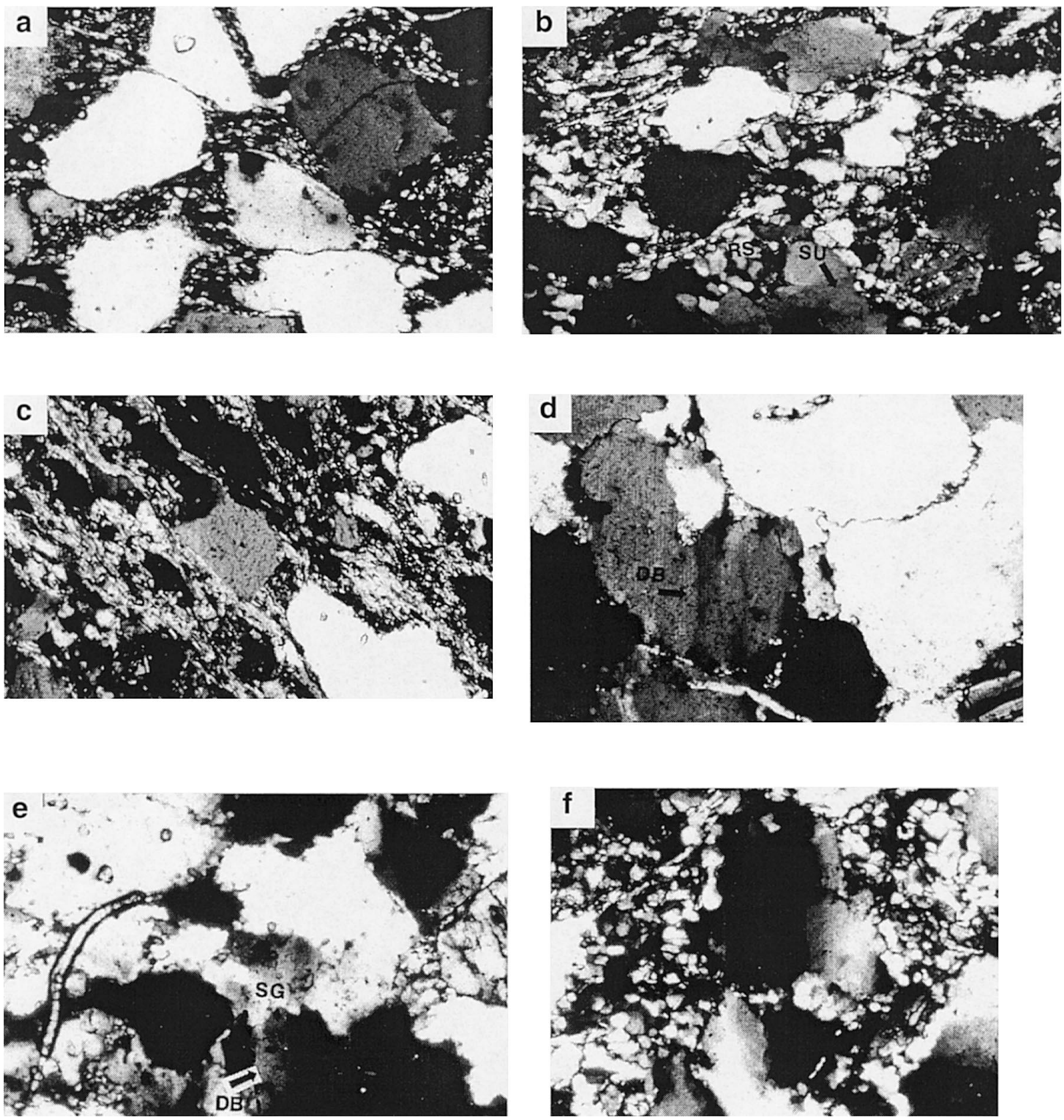

Fig. 10. Microstructural development of rough cleavages and quartz microstructures from the Mt. Bold shear zone. (a) The upper transitional zone to the shear zone comprises less modified quartz grains with mostly low intensity of intracrystalline deformation and weakly developed syndeformational white mica defining the rough cleavage. (b) Within the shear zone there is development of a discontinuous rough cleavage, an increase in the amount of intracrystalline deformation features in quartz grains and a decrease in the size of the quartz grains. Note also the development of subgrains 'SU' and small, recrystallised grains 'RS' in the detrital grains. (c) Continuous rough cleavage occurs along the lower boundary thrust. Syndeformational white micas are more frequently developed forming cleavage zones. Note orientation of detrital muscovite, which shows mica fish-like structure. (d) Slightly deformed quartzite from the upper transitional zone with weak intracrystalline deformation and grain boundary bulge. (e) More intensely deformed quartzite from within the shear zone showing greater development of intracrystalline features like deformation bands (DB) and subgrains (SG). (f) At the basal thrust boundary more highly deformed quartz displays widespread development of recrystallised small grains at the margin of original grains. All thin sections are cut parallel to the stretching lineation and perpendicular to the cleavage and all are crossed polars views with width of view = 3 mm.

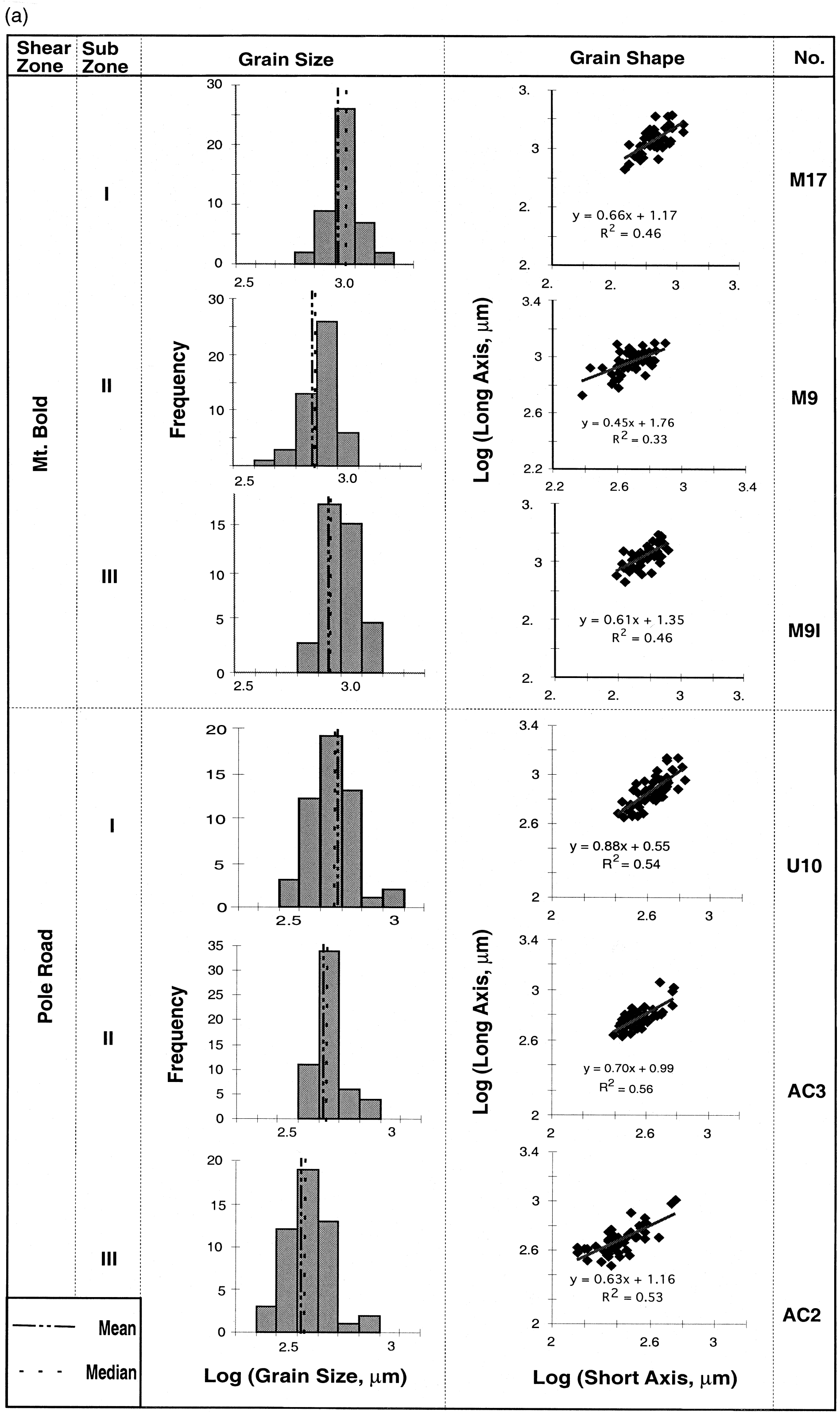

Fig. 11. (a) Distributions of grain sizes and grain shapes of the quartzite from within the Mt. Bold and Pole Road shear zones. (b) Distribution of grain sizes and grain shapes of the quartzite from within the Norton Summit, Greenhill, and Morialta shear zones. Note decrease in the mean value of grain sizes from the transitional zones (sub-zone I) to close to the basal thrusts (sub-zone III) in most shear zones.

(b)

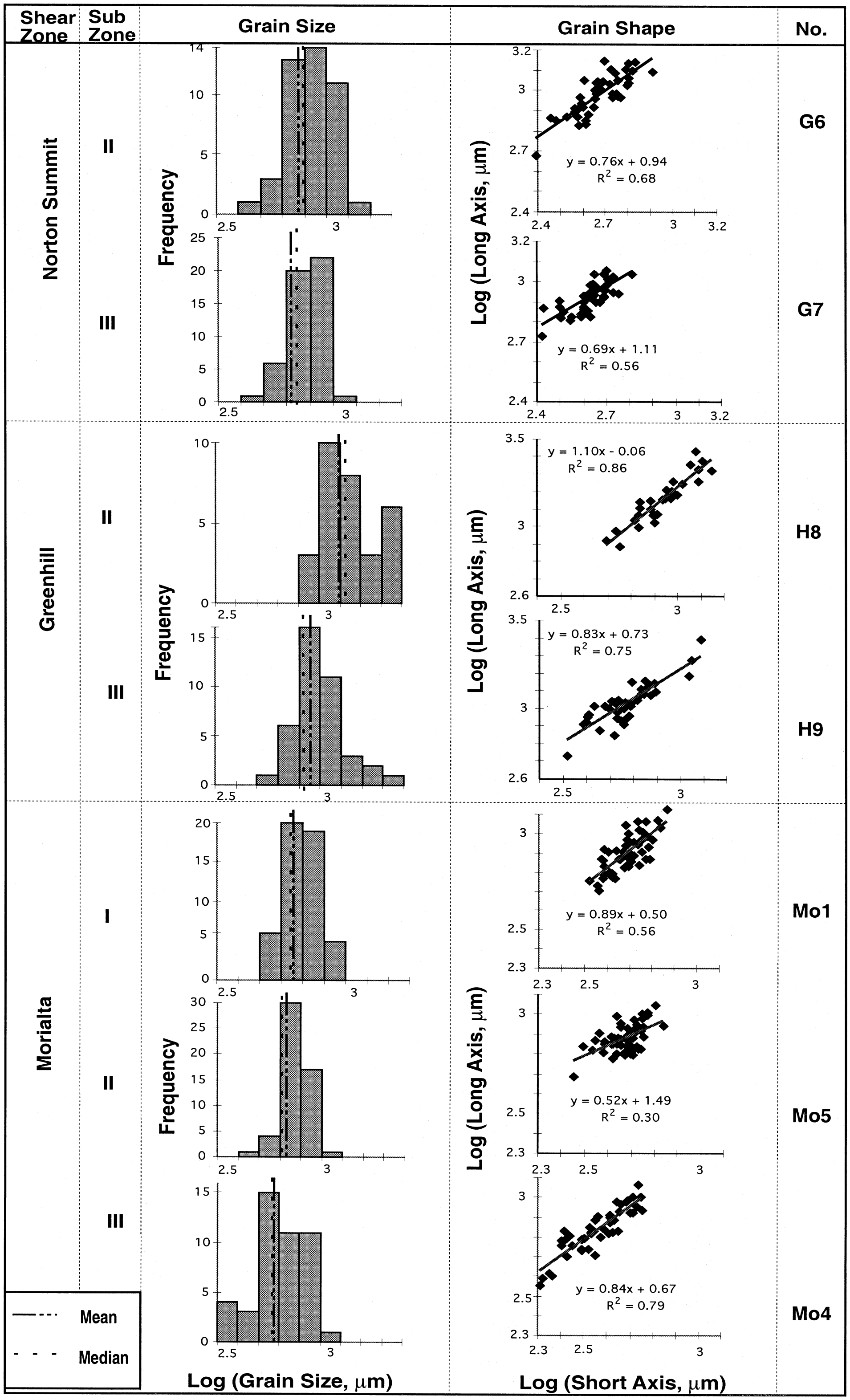

Fig. 11 (*continued*)

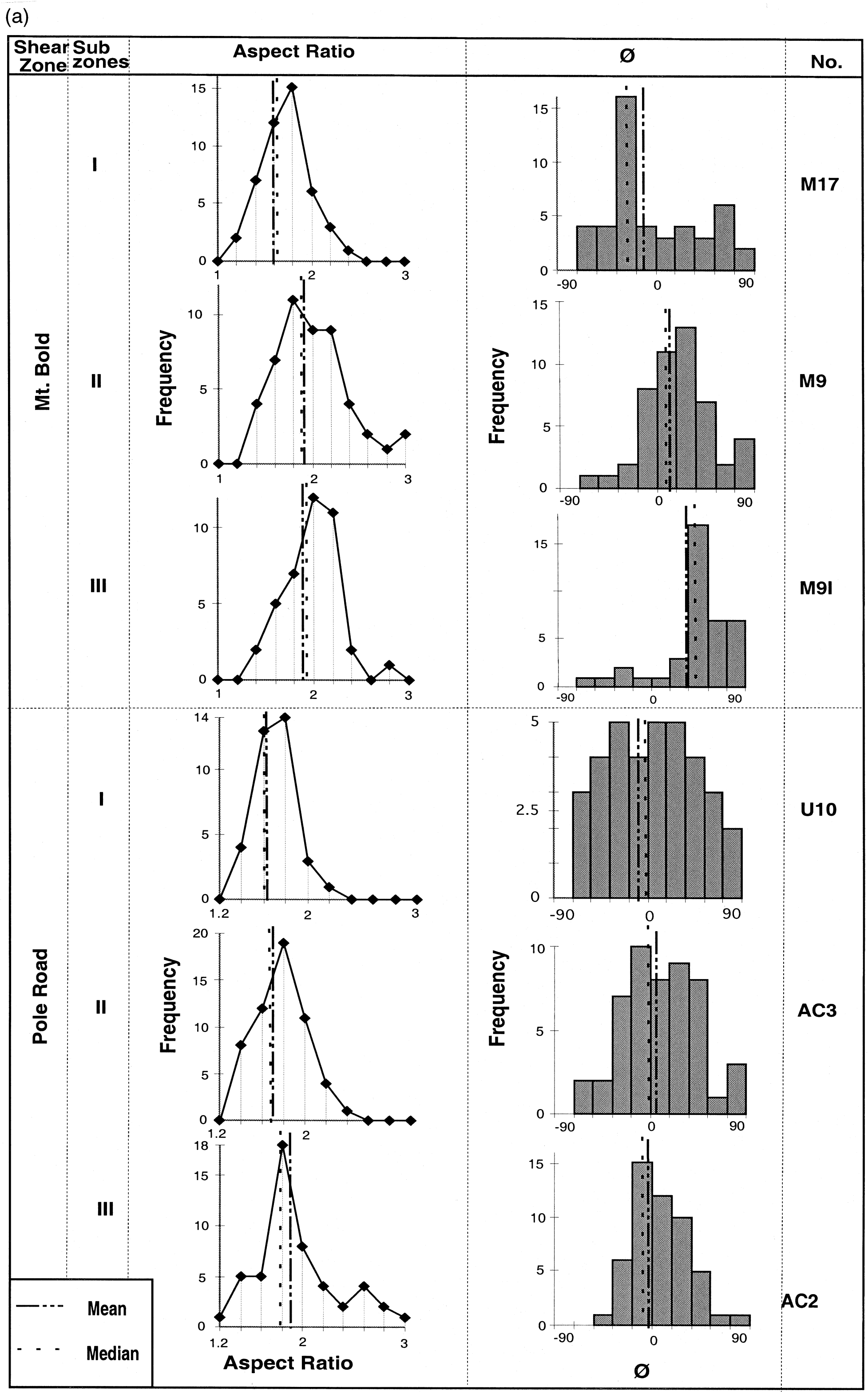

Fig. 12. Aspect ratio and shape-preferred orientations of the grains within quartzite from (a) the Mt. Bold and Pole Road shear zones and (b) Norton Summit, Greenhill, and Morialta shear zones.

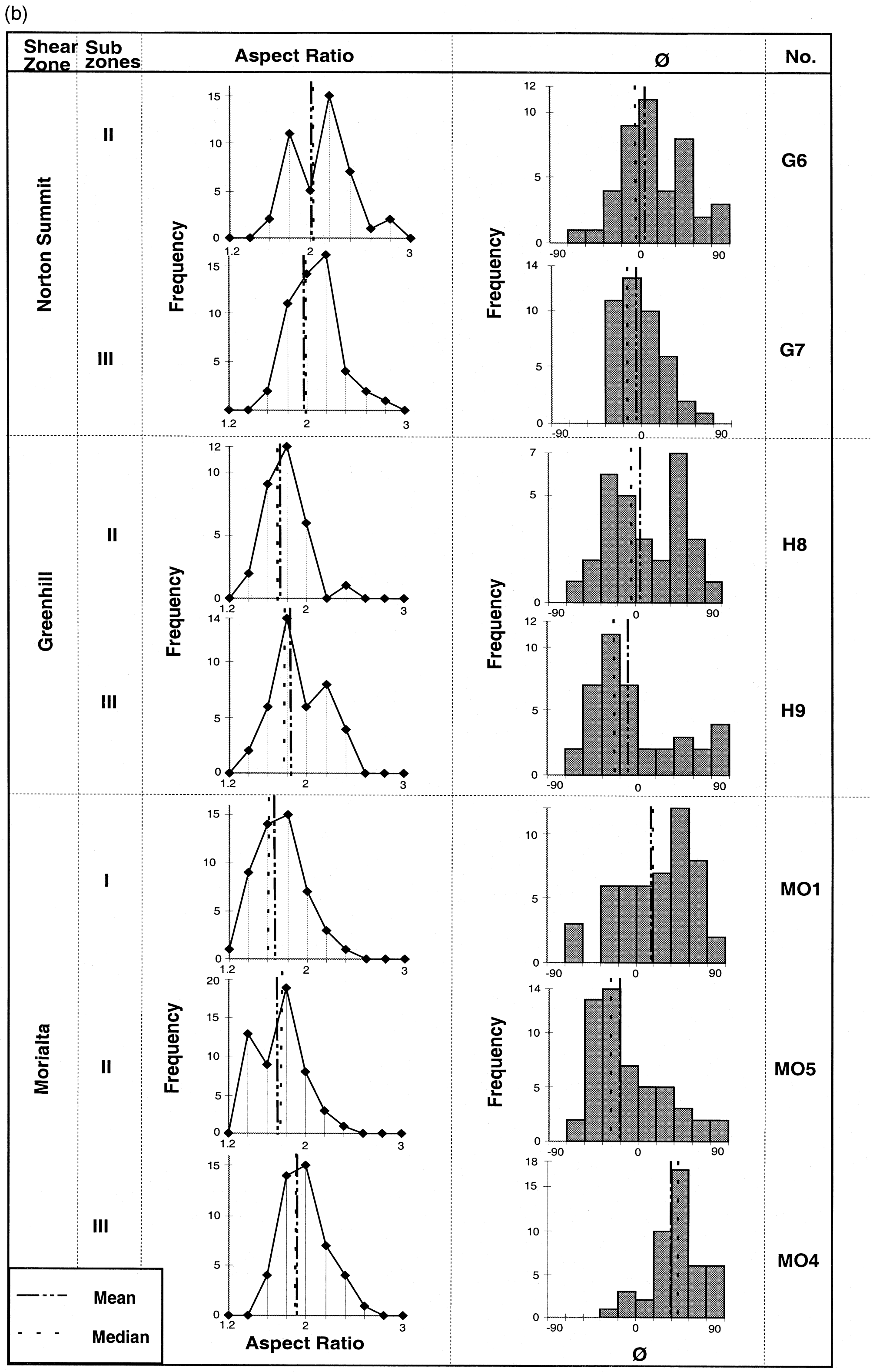

Fig. 12 (*continued*)

finite strain analyses and microstructural studies of the deformational behaviour of the quartz and feldspar.

5.4. Grain-size analysis

The histograms of grain-size variation (Fig. 11) show the distributions of the values of mean and median grain size across the shear zones. In almost all the shear zones the grain size decreases from the upper transitional zones toward the lower boundary thrusts. Scatter graphs of long axis vs. short axis length were also constructed and their regression lines and R^2 values were calculated (Fig. 11). As can be seen on Fig. 11, across all shear zones from the upper transitional zones toward the lower boundary thrusts, the larger grains appear to undergo greater grain size reduction than the smaller ones. Although the values of log-long axis (L) and log-short axis (W) of most of the larger grains is reduced (by about 10% in size), from the upper transitional zones toward the lower boundary thrusts, the dimensions of the smaller grains remain constant. At the lower boundary thrusts, however, the size of the smaller grains is also decreased (by about 15% in size).

Similarly, the range of distribution of the values of long axis vs. short axis lengths from the upper transitional zones toward the lower boundary thrusts across most of the shear zones varies systematically. This can be demonstrated by a decrease in the value of the slope of the regression line and an increase in the y-intercept of the regression line (Fig. 11). As is shown in Fig. 11, in the Pole Road shear zone, the slope of the regression line decreases from 0.9 in the upper transitional zone, to 0.6 at the lower boundary thrust. The decrease in the slope of the trend lines implies that the values of short axes decrease more than the values of the long axes. This suggests that towards the lower boundary thrusts the grains are more flattened due to accommodation of tectonic flattening strains at these boundaries.

5.5. Grain shape analysis

The grain shape distributions (Fig. 12) show that the aspect ratio of quartz grains increases from 1.6 in the upper transitional zone to 2.0 toward the lower boundary thrusts. Studies of the other examples across the shear zones shows that this average is almost invariant.

5.6. Shape preferred orientation analysis

The histograms of Fig. 12 display the distribution of the values of ϕ (the angle between long axes of quartz grains and the reference frame, which was typically bedding) across the shear zones. In almost all cases there is a variation in the degree of shape-preferred orientation of the grains across the shear zones. Similarly, long axis orientation of the grains in most samples shows a symmetrical distribution about the plane of the tectonic fabric. However, samples with a weak long axis orientation about the plane of tectonic fabric may be interpreted as (a) representing a pre-existing sedimentary fabric (sample # M17) or (b) recording a slightly greater intensity of deformation (possibly sample # M9I).

6. Discussion

6.1. Development of shear zones

The presence of asymmetric shear zones in the Adelaide Hills has been recognised by the progressive development of a variety of minor structures found within the shear zones, the microstructural analysis of fabric development and grain configuration, and the spatial distribution of strain variations within the shear zones. Fabrics and structures are incipiently developed at the upper transitional zones to the shear zones and developed in greater concentrations and intensities within the shear zones, finally intensifying along the lower boundary thrusts, which make up the base of each of the shear zones.

One example of this progressive evolution is the development of superposed and transposed cleavages from S_{1a} to S_{1b} and finally to S_{1c} within the shear zones (Figs. 6d and 7). This progressive development of cleavages appears to be uniform through all the shear zones. Since these cleavages show groups of surfaces where their components share almost similar orientations, it is preferred to refer to them as a 'composite foliation' (Tobisch and Patterson, 1988). The processes generating this composite foliation in shear zones can involve complex rotation of early foliations by passing through a crenulation cleavage phase, accompanied by transposition of layers or pre-existing cleavages with differentiation due to solution transfer (Tobisch and Patterson, 1988). It had been previously considered (Offler and Fleming, 1968) that these crenulations formed during regional overprinting. However, the presence of crenulation folds and cleavages in all of the shear zones, whilst not external to them, together with the recognition of transposition of bedding and early cleavages (S_{1a}), supports their development as composite foliations during the processes of shearing and thrusting.

Shear bands are another common geometrical element in the area, and it is now commonly believed that shear bands develop uniquely within shear zones (White, 1979; Platt and Vissers, 1980; Passchier, 1991). Shear bands normally form oblique to the trend of the margin of the shear zones in which they are generated and can be used to deduce the sense of shear (e.g. Platt and Vissers, 1980; Platt, 1984; Hanmer and Passchier, 1991; Passchier, 1991; Stock, 1992). The orientation of the shear bands or ECC-fabrics shows that the sense of shear is consistently toward the northwest (Fig. 5a). Since the orientation of the shear bands is not directly related to the axes of finite strain (unlike an *S* surface), i.e. it is oriented at about 15–25° to the bulk simple shear direction (e.g. Platt, 1984), it is considered that these structures are generated in the later stages of deformation (Platt, 1984; Passchier, 1991). Two reasons are cited for this relationship: (1) they do not propagate parallel to the direction of flow and (2) the strain deviates from simple shear because of superposition of coaxial stretching components on simple shear components at the zone of failure (Platt, 1984). This implies that the kinematics of flow in the Adelaide Hills shear zones, containing shear bands, departs from ideal progressive inhomogeneous simple shear.

On the other hand, the geometry of the two fold types and the non-coaxial component of the deformation that affected the shear zones, suggest that a considerable part of the fold development resulted from either passive rotation of pre-existing folds during non-coaxial flow (Sanderson, 1973; Escher and Watterson, 1974), or as a consequence of the passive amplification of small irregularities in the bedding plane, or due to flow during intense non-coaxial deformation, with shear strains >10 (Cobbold and Quinquis, 1980). Although no extensive evidence of intense non-coaxial deformation is seen in the shear zones, evidence for a distributed non-coaxial progressive deformation operating during at least the latter stage of deformation is provided by the presence of sheath-like folds of Type I geometry (isoclinal recumbent folds, Fig. 8b) and by the existence of oblique folds of Type II geometry (asymmetrical folds Fig. 8a). It seems that the folds initiated as west or northwestward-facing inclined to overturned structures with more intense strains on the overturned limbs and with their axes orthogonal to the movement direction. The folds presumably developed as buckle folds within the thrust and shear zones. The very similar geometry of these minor folds to the geometry of fault-propagation folds may imply that the folds developed in this fashion. Therefore it is likely that many of the minor and mesoscopic folds in the shear zones and associated thrusts were developed as consequence of the accommodation of the folds during non-planar thrusting. This is further supported by the absence of downward or eastward facing relationships of these folds and the consistency of the vergence direction of the folds to the W-NW (see Fig. 8a). The fact that in almost all the shear zones the minor folds are oblique to, or subparallel to, the stretching lineations implies that the folds in the shear zones were rotated after their initiation and during shearing (Sanderson, 1973; Escher and Watterson, 1974).

6.2. Shear zone asymmetry and localisation of deformation along the lower boundary thrusts

Quartzite units near the lower boundary thrusts appear to have suffered higher strain magnitudes than those from the upper transitional zones. This accords with the increase in the concentration of deformational fabrics and structures at the lower boundary thrusts and suggests that strain was mainly localised at these boundaries. The greater relative decrease in the mean value of short axes of the quartz grains than the elongation of the long axes at these boundaries further supports for this argument, though these data are not compelling.

A greater development of continuous rough cleavage, concomitant with the simultaneous development of intracrystalline deformation, at the lower boundary thrusts suggests that strain was partitioned between pressure solution and intracrystalline deformation during cleavage formation. This is most likely due to localisation of deformation and/or strain at these boundaries. Such localisation of deformation is considered to occur due to the process of thrust zone softening described by Wojtal and Mitra (1988). The increase in the proportion of recrystallized small grains on the margins of original quartz grains at the lower boundary thrusts leads to marked decrease in the average size of the quartz grains at these boundaries. Similar strain softening mechanisms involving either enhanced porosity due to decrease in flow stress and increased dissolution, or fluid overpressure to enhance microfracture propagation, have been described by Hippert (1998). Hippert and Hongn (1998) further related softening in protomylonite to mylonite transition zones to abrupt changes in microstructure and deformation mechanism, as well as additional reaction-softening involving the breakdown of feldspar to mica. Wibberley (1999), however, concluded that such strain-softening reactions could also be tempered by hardening associated with silica release, cementation and reduced porosity. It is considered here that softening mechanisms associated with the change from plastic-dominated to solution-dominated shear (cf. Hippert and Hongn, 1998), lead to the development of core and mantle structures and localisation of deformational features along the lower boundary thrusts.

6.3. Deformation mechanisms operating during development of the shear zones

The deformation mechanisms which operated during development of the shear zones can be inferred from the deformation microstructures observed in quartzites and psammites within the shear zones. Strain, and therefore displacement along the shear zones, was accommodated by crystal-plastic dislocation creep (Knipe, 1989) in quartzites and pressure solution (Beach, 1979; Rutter, 1983) in psammites. Observations on the deformational behaviour of quartz and feldspar grains in the quartzite units from within the shear zones show that quartz grains exhibit mainly intracrystalline deformation features while feldspar grains show mainly internal fracturing. However, locally quartz grains show evidence of dynamic recrystallisation. The intensity of intracrystalline deformation features, like sweeping undulatory extinction and deformation bands in quartz grains increases from the upper transitional zones towards the lower boundary thrusts. Along the lower boundary thrusts, softening mechanisms reduced the work-hardening process and increased the ductility of the quartzite to produce the core and mantle structures (Fig. 10f) (White et al., 1980; Schmid, 1982).

In contrast to the quartzite, observations of the microstructural development of cleavages in psammites from within the shear zones reveals that the cleavages developed by pressure solution processes. The existence of insoluble residues in foliae or solution seams, sutured quartz grain boundaries, truncation of detrital quartz grains against foliae, which comprise insoluble phyllosilicate grains (Fig. 8h), and offset across foliae, together with enhancement in the development of rough cleavage from the discontinuous type to the continuous type from within the shear zones, all demonstrate that pressure solution was an important, if not the dominant mechanism of cleavage formation. However, the abundant evidence of intracrystalline deformation in quartz grains such as deformation bands, sweeping undulatory extinction and occasional subgrain formation (Fig. 10b), from the lower boundary thrusts, implies that intracrystalline deformation has also contributed to the development of the rough cleavages.

6.4. A structural model for the progressive development of the asymmetric shear zones and associated imbricate thrusts

The spatial concentration of fabrics and structures in the shear zones and their associated imbricate thrusts suggests that the deformation in the Adelaide Hills area and in general in the SAF–TB is heterogeneous on a large scale and is partitioned into zones of high-strain ductile deformation commonly sub-parallel to the bedding (S_0) and/or regional cleavage (S_{1a}). The near parallelism of S_{1a} within and outside the shear zones and associated thrusts, the consistency of vergence direction of regional and fault-related cleavages, the roughly constant orientation of the stretching lineations and the preservation of minor, asymmetric shear-sense criterion suggest the shear zones were dominated by a significant non-coaxial deformation history.

The near parallelism of S_{1a} within and outside the shear zones also suggests that the deformation should not be modelled as a classical heterogeneous simple shear. This is due to the requirement of marked variation in cleavage orientation that parallels the XY plane of finite strain between different domains of strain intensity (Ramsay and Graham, 1970) (Fig. 3a). The presence of shear bands in the shear zones which are recognised to develop in the later stages of deformation during general non-coaxial flow (Platt and Vissers, 1980; Platt, 1984; Hanmer and Passchier, 1991) may further constrain this argument. Furthermore, the extensive evidence of oblique to sub-parallel folds (Type I) along with the presence of recumbent folds (Type II) cross-cut by imbricate thrusts or shear zone branches (e.g. Fig. 8d), also implies that the structures in the area developed in a progressive manner rather than during single, short-lived events.

Similarly, the W-NW overturned character of the folds and the fault-propagation geometry of the major and mesoscopic folds implies a component of overthrusting in this direction. The simplest explanation for the coexisting evidence of non-coaxial deformation and overthrusting is that there was an early phase of W-NW directed layer-parallel shortening and thrusting, possibly due to a decrease in thrust movement near thrust tips (Coward, 1984; Coward et al., 1992). This strain history implies a combination of homogeneous pure shear and heterogeneous simple shear (Sanderson, 1982). This early deformation was progressively overprinted by a deformation which would have caused, for instance, passive rotation of early buckle folds to produced oblique to sub-parallel folds (Hansen, 1971; Escher and Watterson, 1974). This later deformation is formed by layer-parallel extension and shearing which caused the rocks to stretch and flatten sub-parallel to the shear plane (Sanderson, 1982, Passchier, 1991). In the case of the study area, the lower boundary thrusts to the shear zones form the shear planes during localisation of deformation or development of the shear zones. The presence of shear bands, which are considered to have developed in extensional shear zones (Passchier, 1991), may further constrain the progressive superposition of this later stage of shear zone development. Similar strain histories within thrust zones have also been proposed elsewhere, for instance, in the Betic Cordillera of SE Spain (Platt and Behrmann, 1986), in northern Scotland (Holdsworth, 1989, 1990), and in central Australia (Kirschner and Teyssier, 1992).

6.5. Local significance of the structural model

The Adelaide Hills area lies in the northern part of the Southern Adelaide Fold–Thrust Belt adjacent to the foreland to the Cambro/Ordovician Delamerian Orogenic Belt of South Australia (Flottmann et al., 1994, 1995; Flottmann and James, 1997). Within the belt, the Neoproterozoic Adelaidean and Cambrian Normanville/Kanmantoo Groups were reactivated and inverted and their sedimentary rocks were thrust west

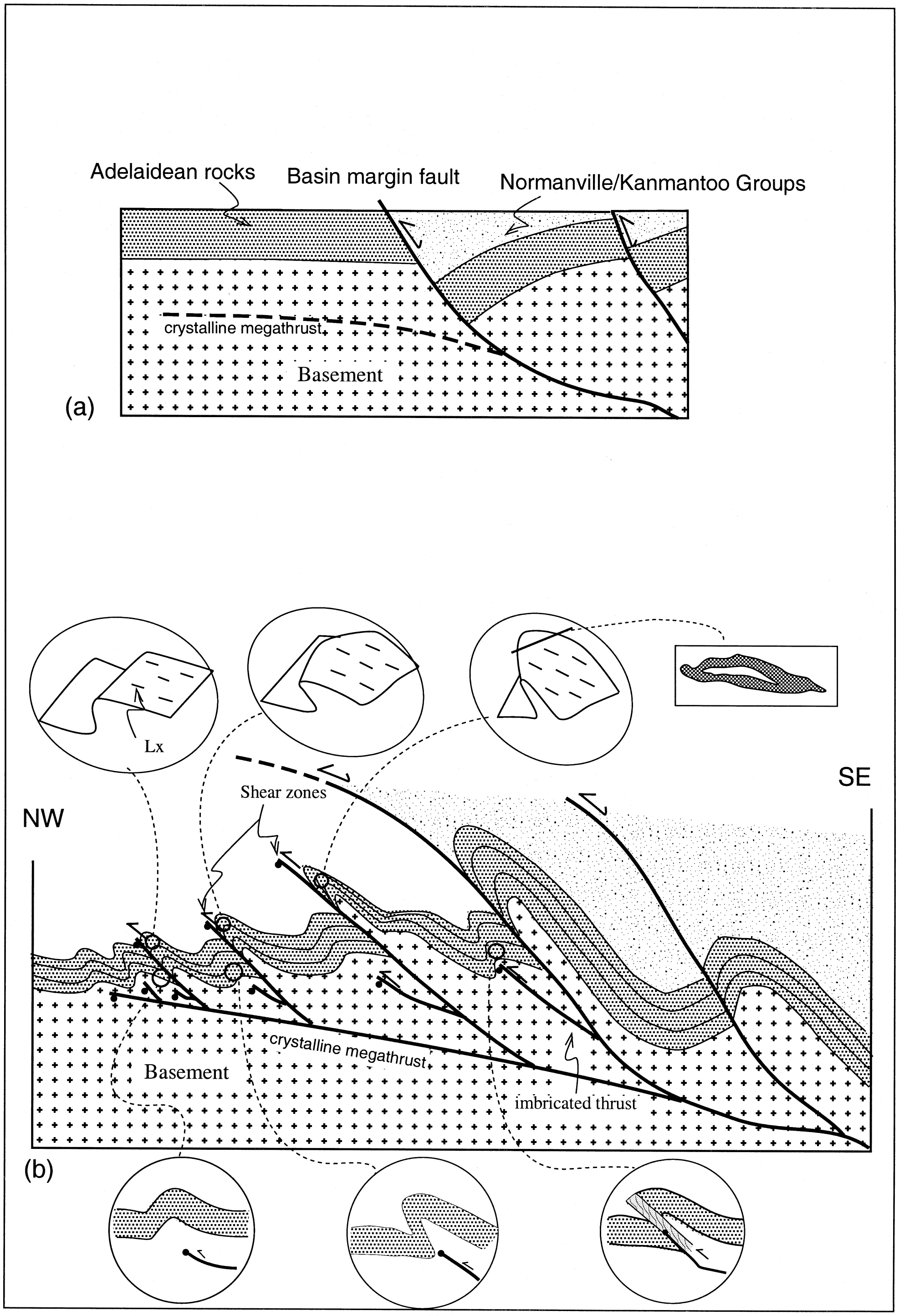

Fig. 13. Schematic diagrams showing development of the shear zones and associated ductile thrusts. (a) Initiation of crystalline megathrust in basement during basin reactivation; (b) Propagation of the thrusts toward the foreland, the strain field of combined layer-parallel stretching and shearing, which caused development of the shear zones, is superposed. For further explanation see text.

toward the Australian Craton (Flottmann and James, 1993, 1997). During this reactivation, a master décollement within the Palaeoproterozoic basement crystalline initiated a megathrust sheet (cf. Hatcher and Hooper, 1992) (Fig. 13a). The propagation of this basal thrust in a W-NW direction toward the foreland was accomplished by shearing of imbricate thrust/shear zones, which caused the development of fold–thrust interaction buckle folds (fault-propagation folds) in the hanging wall of the propagating ductile thrusts (Fig. 13b).

The amount of shearing decreased toward the thrust sheets closer to the foreland and the deformation was more localised along the lower boundaries of the pre-existing thrust sheets. Closer to the hinterland dynamic recovery/recrystallisation aided development of more intense shear zones (Fig. 13b). This is constrained by the decrease in the frequency and intensity of the shear bands (ECC-fabrics) along with the absence of curvilinear folds from the Mt. Bold shear zone toward the Morialta shear zone. The localisation of deformation along the lower boundary thrusts was developed during layer-parallel extension and shearing deformation. Nevertheless, such localisation of deformation due to layer-parallel extension and shearing deformation did not occur along associated imbricate thrusts, which propagated in both the hanging wall and the footwall to the shear zones. Therefore, ductile deformation of these thrusts was more likely developed only due to layer-parallel shortening and thrusting deformation (Fig. 13b). This is also justified by the absence of the shear bands and by the very narrow width of the deformation zones along these imbricate thrusts in comparison with the major shear zones.

7. Conclusions

- The spatial, geometric and kinematic distribution of highly deformed fabrics and structures within the shear zones and their associated imbricate thrusts show that the deformation in the external portion of the SAF–TB is mainly partitioned between narrow highly deformed zones or shear zones, separated by larger less deformed zones with fewer deformation features. Fabrics and structures within these highly deformed shear zones are incipiently developed at the upper transitional zones to the shear zones and then progressively develop more intensely within the shear zones and finally intensify along the lower boundary thrusts, which make up the base to the shear zones. The presence and intensity of these fabrics and structures (for example shear bands) decrease from the shear zones closer to the hinterland towards the shear zones closer to the foreland.
- Macroscopic and microscopic kinematic indicators including shear bands, asymmetric minor folds, asymmetric pressure fringes, and displaced broken grains show that the shear zones were propagated and overthrust consistently towards W-NW.
- Observation of microstructural development of cleavages in psammites shows that cleavages are well developed within the shear zones by the process of pressure solution. Quartzites exhibit overwhelming evidence of crystal-plastic deformation mechanisms across the shear zones. The dominance of this deformation mechanism decreases from the shear zones closer to the hinterland towards the shear zones on the foreland side where evidence of pressure solution deformation mechanisms is also seen in quartzites.

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