Journal of Structural Geology, Vol. 8, Nos. 3/4, pp. 457 to 472, 1986
Printed in Great Britain

0191-8141/86 $03.00 + 0.00

Ductile thrusting within subduction complex rocks on Signy Island, South Orkney Islands

A. W. MENEILLY and B. C. STOREY

British Antarctic Survey, Natural Environment Research Council, High Cross, Madingley Road, Cambridge CB3 0ET, U.K.

(*Received* 17 *June* 1984; *accepted in revised form* 30 *August* 1985)

Abstract—A model involving repeated NNW-directed ductile thrusting in subduction complex rocks on Signy Island, South Orkney Islands, accounts for the consistent NNW-trending alignment of fold axes, stretching and shearing directions. In the absence of major discrete thrust planes, stratigraphic control or younging criteria, the evidence for consistent ductile thrusting is more cryptic and is found in folding/stretching relations, porphyroblast rotations and mylonite fabrics. Early fold axes parallel to stretching, syn-tectonic garnet rotations, augen lineations and sheath folds all support NNW-directed thrusting as do a common late shear-band foliation and minor shear zones. Development of late open folds with axes parallel to stretching may be due to constrictional deformation, a lateral ramp or differential movement during thrusting.

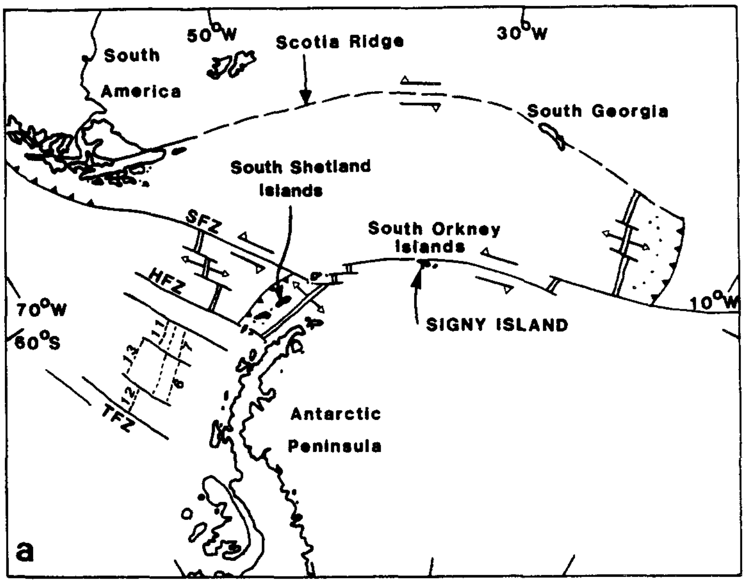
The structures described from Signy Island may characterize the type of deformation in deep-level ductile thrust zones at the base of an accretionary wedge or in a zone of subcretion. The thrusting direction on Signy Island is inferred to be parallel to the direction of subduction in the South Orkney Islands in the early Jurassic: plate reconstructions suggest that this direction was at a high angle to the arcuate trend of the Antarctic Peninsula continental margin.

INTRODUCTION

SIGNY Island (60°43′S, 45°38′W) in the South Orkney Islands lies on the southern Scotia Ridge (Fig. 1). It is part of the Scotia metamorphic complex (Tanner *et al.* 1982), a Mesozoic–(?)Cenozoic accretionary subduction complex formed along the Pacific margin of Gondwanaland (Smellie & Clarkson 1975, Barker & Dalziel *et al.* 1976, Rivano & Cortes 1976, Súarez 1976, de Wit 1977, Smellie 1981, Dalziel 1982). The Scotia metamorphic complex, cropping out on the South Shetland and South Orkney Islands (Fig. 1), consists of interlayered graphitic phyllite, mica–schist, metachert, marble, calc-silicate rock and metavolcanic rocks (Thomson 1968, 1974, Dalziel 1982) and includes a dunite–serpentinite sheet on Gibbs Island (De Wit *et al.* 1977). The metamorphic grade is commonly epidote–amphibolite or amphibolite (Tanner *et al.* 1982) but greenschists and glaucophane–lawsonite-bearing blueschists are present in the South Shetland Islands (Smellie & Clarkson 1975, Rivano & Cortes 1976, Dalziel 1982). The Scotia metamorphic complex on the South Orkney Islands has yielded K–Ar cooling ages on hornblende and micas of about 190 Ma (early Jurassic) (Miller 1960, Grikurov *et al.* 1967, Rex 1976, Tanner *et al.* 1982) and is overlain by Upper Jurassic–Lower Cretaceous conglomerates.

The main lithologies on Signy Island are mica–schist, marble and amphibolite (Matthews & Maling 1967, Thomson 1968). Matthews & Maling (1967) divided the sequence into three 'series', the Moe Island (mica–schist), the marble, and the amphibolite 'series', based on the proportions of the three main rock types. Although the lithologies can be crudely grouped into the above units the complex refolding, tectonic contacts between lithologies and complete absence of younging evidence make stratigraphical correlations difficult. The mica–schists comprise mappable units of pelitic or semi-pelitic schist, psammite and rare quartzite layers. The amphibolites include fine-grained greenschist, coarse hornblende gneiss, epidote amphibolite and calc–amphibolite. Massive white calcite–marble, in bands up to 3 m thick, is the most distinctive rock on Signy Island. Marble is usually associated with amphibolite, commonly interlayered on a centimetre scale and is also found in laminated psammite–quartzite–amphibolite–marble units. Whole-rock chemical analyses by the authors (Storey & Meneilly 1985) suggest that the Signy Island amphibolites are metamorphosed basaltic rocks and have affinities with MORB-type and alkali basalts of an intraplate oceanic island tectonic setting. Finely laminated, thin, Mn-rich garnetiferous quartzite layers are interpreted as meta-chert. The protoliths on Signy Island (ocean basalt, limestone, chert, clastic sediments) are typical of subduction zone environments (Kanehira 1967, Dickinson 1971, Wood 1978, Jacobson 1983).

The structure of Signy Island has been described briefly by Matthews & Maling (1967) and Thomson (1968). A more detailed account of the structure of the Scotia metamorphic complex is given by Dalziel (1984) who emphasized the intense and pervasive ductile shearing, tectonic interleaving of lithologies and coaxial folding. He suggested that the deformation and accompanying metamorphism may represent the subcretion (Karig & Kay 1981) of material from a downgoing oceanic plate beneath the fore-arc wedge at depths in excess of 20 km in the subduction zone. Although recognizing the role of intense shearing, Dalziel (1984) did not determine the direction of tectonic transport involved. Evidence is



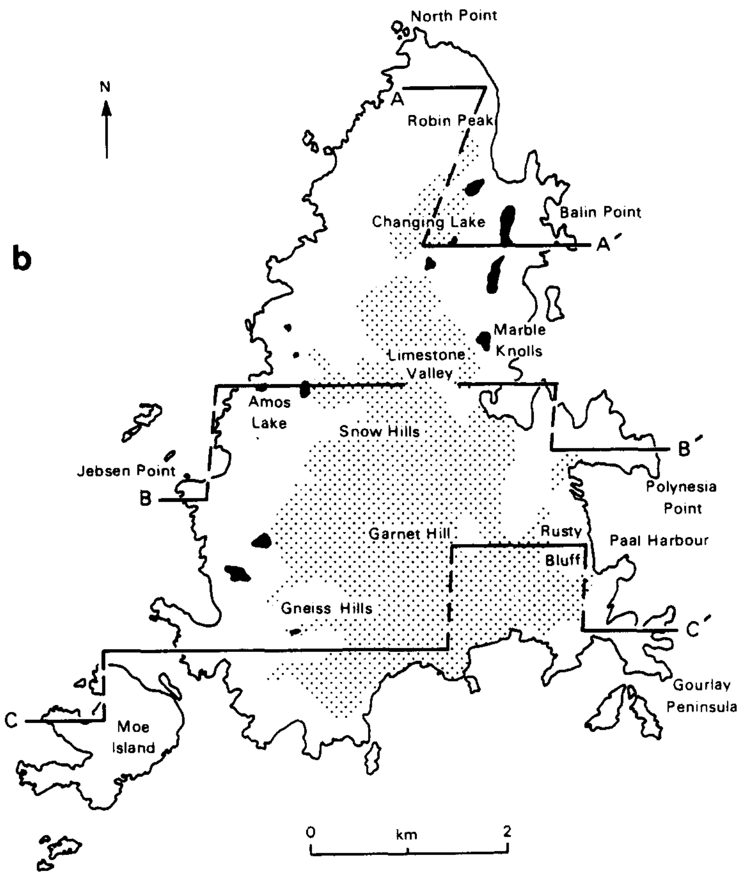


Fig. 1. Map of the Scotia arc region and location map of Signy Island showing lines of sections in Fig. 2. Present active subduction zones, spreading ridges and transform faults and numbered magnetic anomalies after Barker (1982). SFZ, Shackleton fracture zone; HFZ, Hero fracture zone; TFZ, Tula fracture zone. Permanent ice is stippled, lakes are in black.

presented below to show that the NNW–SSE axial trend of folds on Signy Island is sub-parallel to the direction of tectonic transport determined from stretching lineations, porphyroblast rotation and mylonite fabrics.

DEFORMATION D_1–D_5

The major structures and the relationships of the main fold phases (F_2–F_5) are illustrated by cross-sections (Figs. 2 and 3). The age relationship of mineral growth to deformation events is shown in Fig. 4.

(?) D_1 *and* D_2

The earliest lithological layering on Signy Island, which includes major units of amphibolite, mica–schist or marble more than 100 m thick and fine felsic- and mafic-rich laminations, is designated S_1. S_1 may repre-

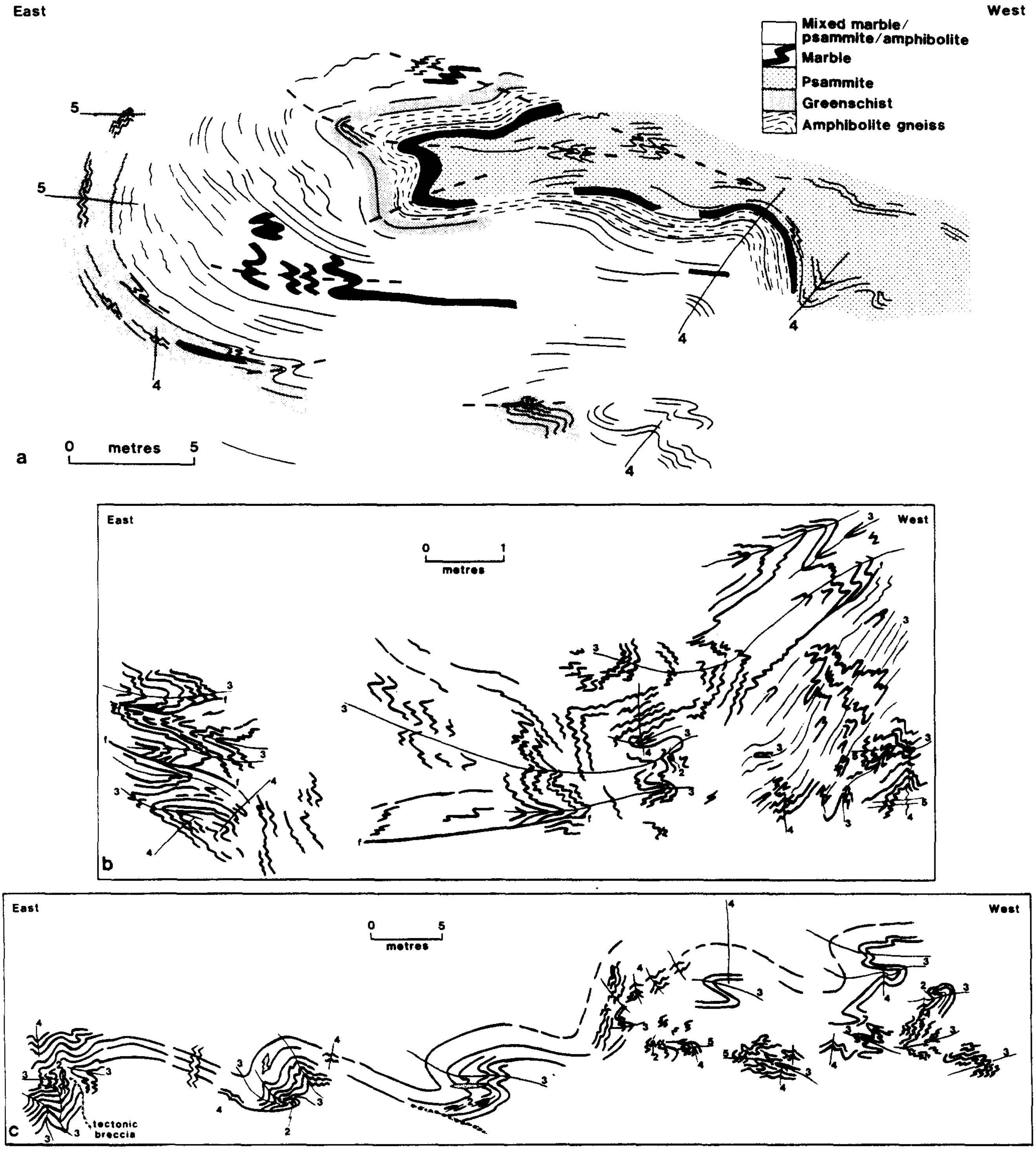


Fig. 3. Fold relations drawn from field sketches and photographs. (a) Large F_3 fold 500 m SE of North Point. Heavy dashed lines are F_3 fold axial planes. (b) All four phases of folding seen on the coast north of Balin Point. Heavy lines are lithological banding (S_1); fine lines are fold axial planes or cleavage. f, syn-D_3 faults. (c) Section along coast at Jebsen Point. Stipple, quartz segregations.

sent D_1 metamorphic segregations but no D_1 folds, lineations or mineral preferred orientations were recognized. No sedimentary structures were found and it is not known if lithological layering represents bedding, transposed bedding or tectonic layering. Alternating 10 cm layers of marble, mica–schist and amphibolite, however, may represent original bedding (possibly with concordant basalt sheets).

Generally S_2 is parallel to S_1 resulting in a composite $S_{1/2}$ foliation but S_2 is axial planar to small-scale isoclinal and intrafolial folds of S_1. S_2 is affected by several fold phases but generally it has a shallow dip to the south (Fig. 5a). Complete recrystallization or neomineralization of mica, amphibole, quartz, plagioclase, epidote and ore minerals took place syn- and/or post-D_2 (Fig. 4). In F_2 fold hinges S_2 is a pervasive dimensional and crystallographic preferred orientation of all minerals. Any D_1 mineral fabric which may have existed has been

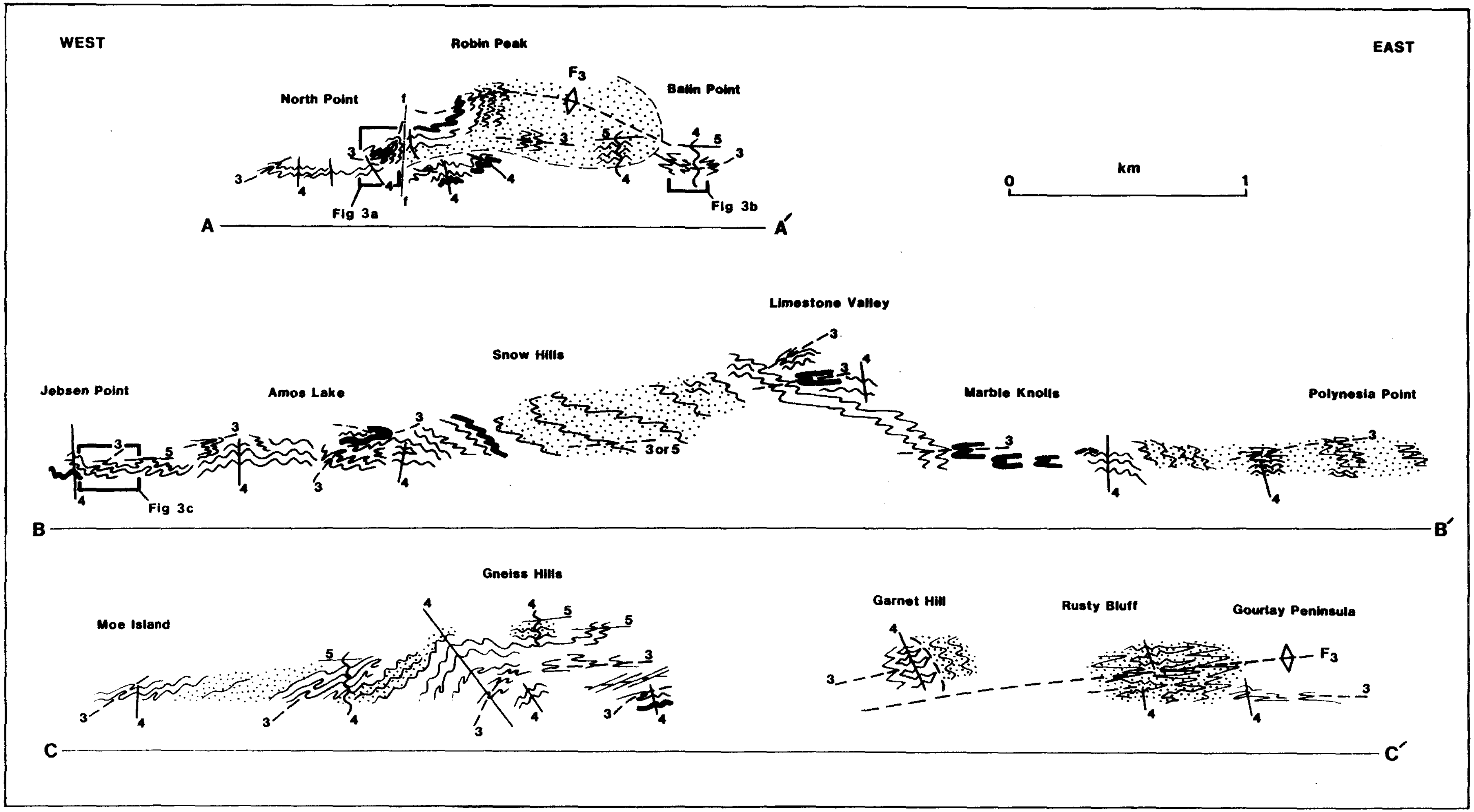


Fig. 2. Schematic, composite cross-sections showing major structures and relationships of F_3, F_4 and F_5; numbers refer to deformation phases and lines of section are shown on Fig. 1. Thick psammite units are stippled and marble bands are in solid black.

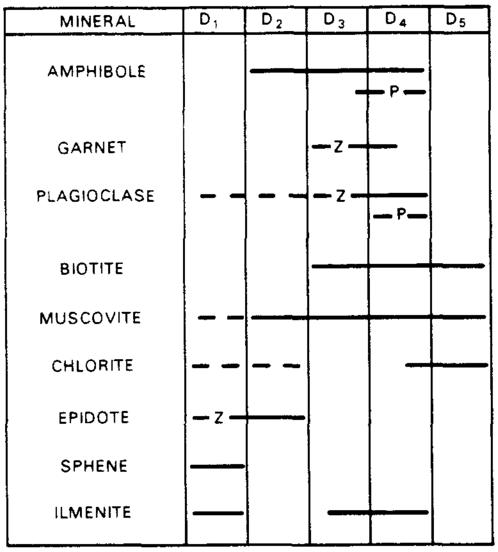


Fig. 4. Age relationship of mineral growth to deformation events. Dashed line indicates inferred growth; P refers to growth of amphibole and plagioclase porphyroblasts, Z refers to zoned minerals.

completely destroyed and S_1 cleavage is now defined only by small-scale felsic- and mafic-rich layers or by inclusion trails of fine opaque material in syn-D_2 plagioclase porphyroblasts. Hinge-lines of intrafolial F_2 folds and long axes of ductile boudinage of S_1 produce a well-developed L_2 rodding lineation (Fig. 6a); this generally plunges SSE but is locally variable with a SW trend common in some areas (Fig. 5a). In finely banded mica-schist and greenschist D_2 produces characteristic dismembered S_1 quartz–feldspar layers with small isolated F_2 fold noses (Fig. 6b). A fine, syn-D_2 mineral elongation-lineation (X_2), which probably represents the D_2 stretching direction, also plunges generally SSE parallel to L_2 and is folded by F_3 and later folds. Where L_2 has a variable trend the mineral lineation remains SSE trending but recognition of the X_2 stretching lineation is complicated by later SSE-trending F_3 or F_4 folds with their own hinge-parallel mineral lineations. No consistent vergence (Roberts 1974) or change of vergence of F_2 folds could be mapped and no major F_2 folds were recognized.

D_3

Open to tight F_3 folds plunging gently SSE with shallow dipping axial planes and short limbs of 10–100 cm are common throughout Signy Island (Figs. 5a and 6b–d). Mapping of the vergence of F_3 folds reveals a major F_3 antiform with axial trace trending NW–SE through the centre of the island (Fig. 5a). Prominent features of this major structure are:

(1) the transposition of $S_{1/2}$ foliation in psammites by tight F_3 folds in the hinge region around Paal Harbour,

(2) east vergence of F_3 folds in the SW part of the island and on Moe Island,

(3) west vergence of F_3 folds in the north and NW part of the island.

Intense D_3 deformation occurs locally with isoclinal folding, shearing out of fold limbs, syn-D_3 faults (Fig. 6d) and rare sheath folds with axes aligned NNW–SSE. Syn-D_3 garnet porphyroblasts have quartz-rich pressure shadows and an augen lineation trending NNW–SSE. Thin sections show that the axis of rotation of syn-D_3 garnet porphyroblasts was approximately normal to the F_3 fold axes and rotations of up to 360° occurred (Fig. 7a). During simple shear a rigid spherical porphyroblast in a deformable matrix will rotate according to $\omega = \gamma/2$, where ω is the angle of rotation and γ the shear strain. Thus, a rotation of 360° requires a minimum D_3 shear strain of 12.5. Of ten thin sections examined, garnet porphyroblasts in eight showed a sense of rotation consistent with NNW-directed thrusting parallel to foliation. Two uncommon but distinctive rocks are developed during this D_3 shearing and occur in thin bands on the west side of the island. One consists of small lenticles of muscovite–schist or quartz bounded by discrete shears and set in a matrix of fine hornblende schist (Fig. 7b); it has a strong rodding and mineral lineation plunging SSE. The other is a small-scale tectonic breccia of dismembered thin quartz layers in a micaceous matrix. Minor syn- or post-D_3, pre-D_4 faults striking N–S are common.

Syn-D_3 garnet porphyroblasts contain inclusions of pre- and syn-D_3 albite porphyroblasts. Where D_3 deformation is intense, recrystallization of quartz, albite, mica and amphibole produce a pervasive S_3 schistosity in F_3 hinge zones. Large hornblende porphyroblasts are late- or post-D_3. They usually have their long axes parallel to S_2 and on S_2 surfaces may be randomly oriented, in rosettes, or have a strong alignment parallel to the dominant SSE lineation.

D_4

F_4 folds are generally less tight and less common than F_3 folds (Figs. 6b & c). They plunge parallel to F_3 axes and their axial planes are upright or east dipping (Fig. 5b). Upright F_4 folds with wavelengths of 1–5 m are the most common folds in some areas particularly in the north and west of the island (Figs. 5b and 2). On the lower limb of the major F_3 antiform, F_3 and F_4 folds both verge to the west and the two can be difficult to distinguish. Garnet porphyroblasts with tight S_4 crenulations wrapping around them and containing open S_4 crenulations of S_2 trails are interpreted as syn-D_4. Subhedral garnet overgrowths on syn-D_3 garnet are probably also syn-D_4. Rotated inclusion trails in syn-D_4 garnet imply a northerly vorticity during growth. Quartz, plagioclase, hornblende, chlorite, mica and ore are recrystallized locally during D_4, producing a coarse S_4 schistosity.

D_5

F_3 and F_4 folds are cut by F_5 folds with flat-lying axial surfaces, less common than the two earlier phases but with a similar SSE axial trend (Fig. 5b). F_5 folds are commonly associated with narrow shear zones on one limb implying overthrusting in a westerly direction. A common shear band foliation dipping gently to the NW is probably also of D_5 age and is described in detail below.

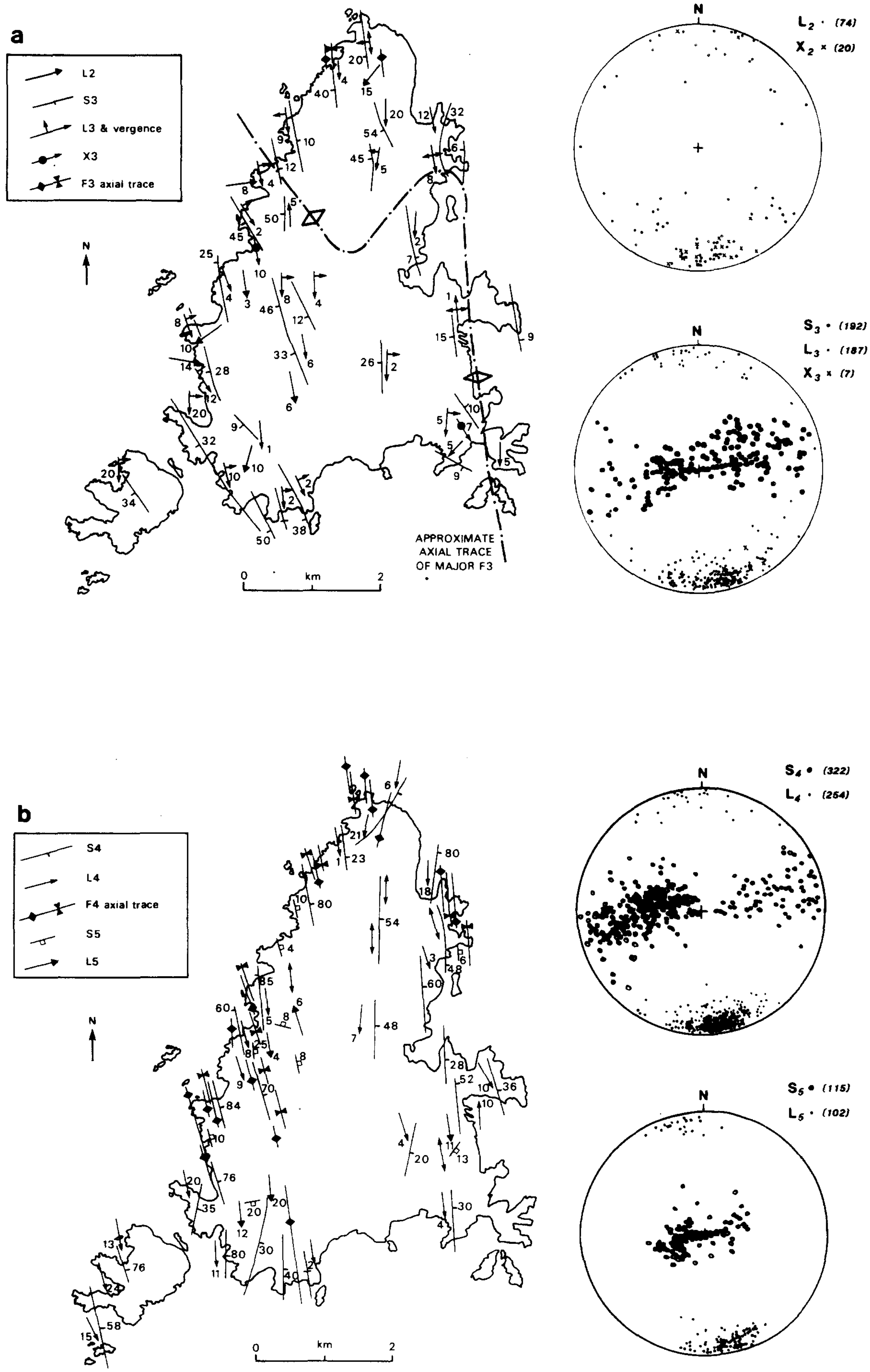


Fig. 5. Maps and equal area stereograms of (a) D_2/D_3 and (b) D_4/D_5 structures on Signy Island. L denotes lineation of hinge-line and X the elongation lineation.

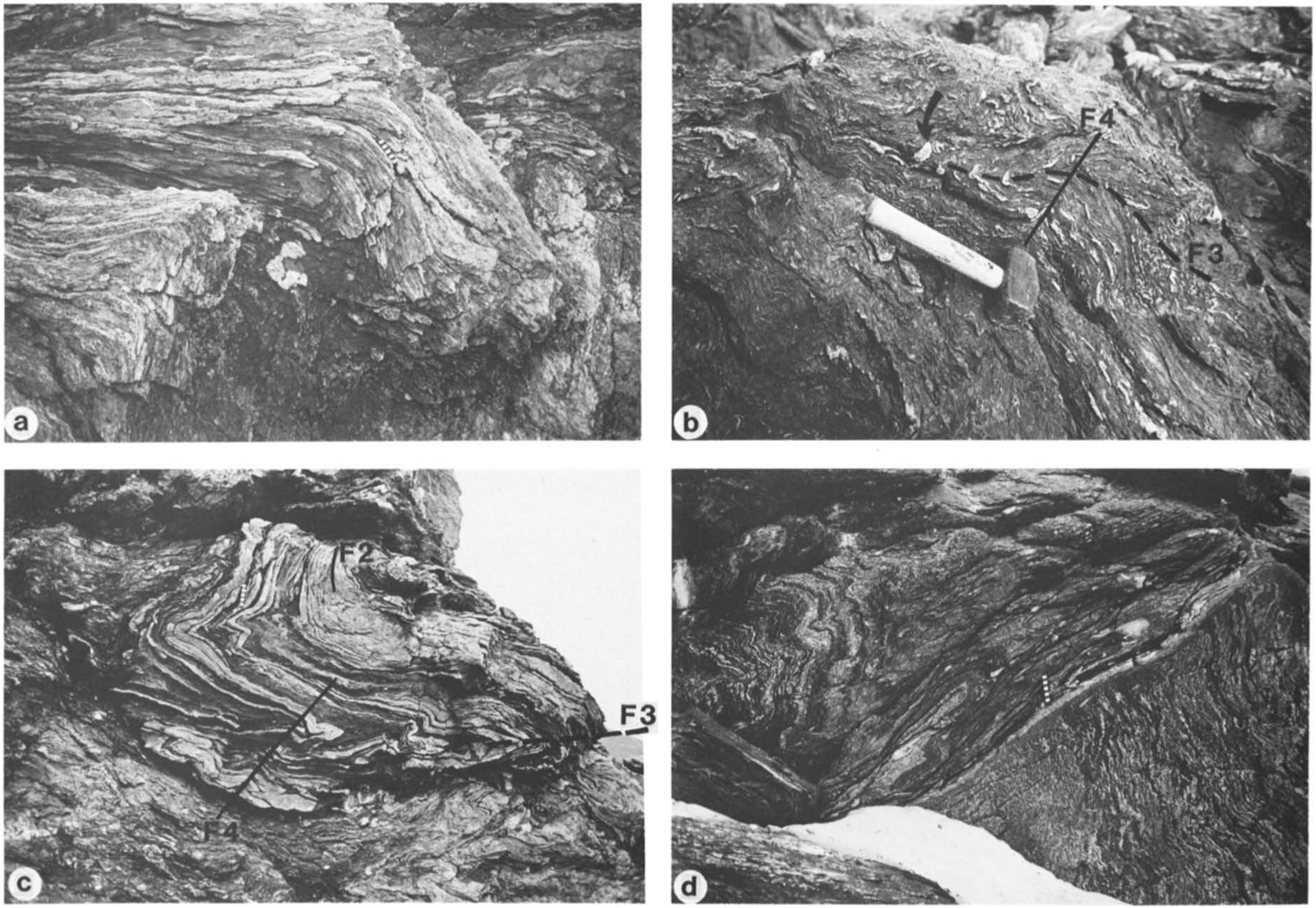


Fig. 6. (a) L_2 rodding lineation folded by F_4 folds near Jebsen Point. The scale is 10 cm long. (b) F_3 folds in greenschist at Changing Lake with axial plane dipping west folding S_1 which is dismembered by D_2 shearing and intrafolial folding (arrow indicates F_2 fold). Open F_4 folds with axial planes dipping east refold F_3. Hammer handle is 20 cm long. (c) F_3 folds with horizontal axial planes in banded psammite–amphibolite from near Jebsen Point. Small F_4 folds have axial planes dipping east. Note the intrafolial F_2 fold in psammite at top of photograph. Scale is 10 cm long. (d) F_3 folds intensifying towards fault plane. West is to the right and the scale is 10 cm long.

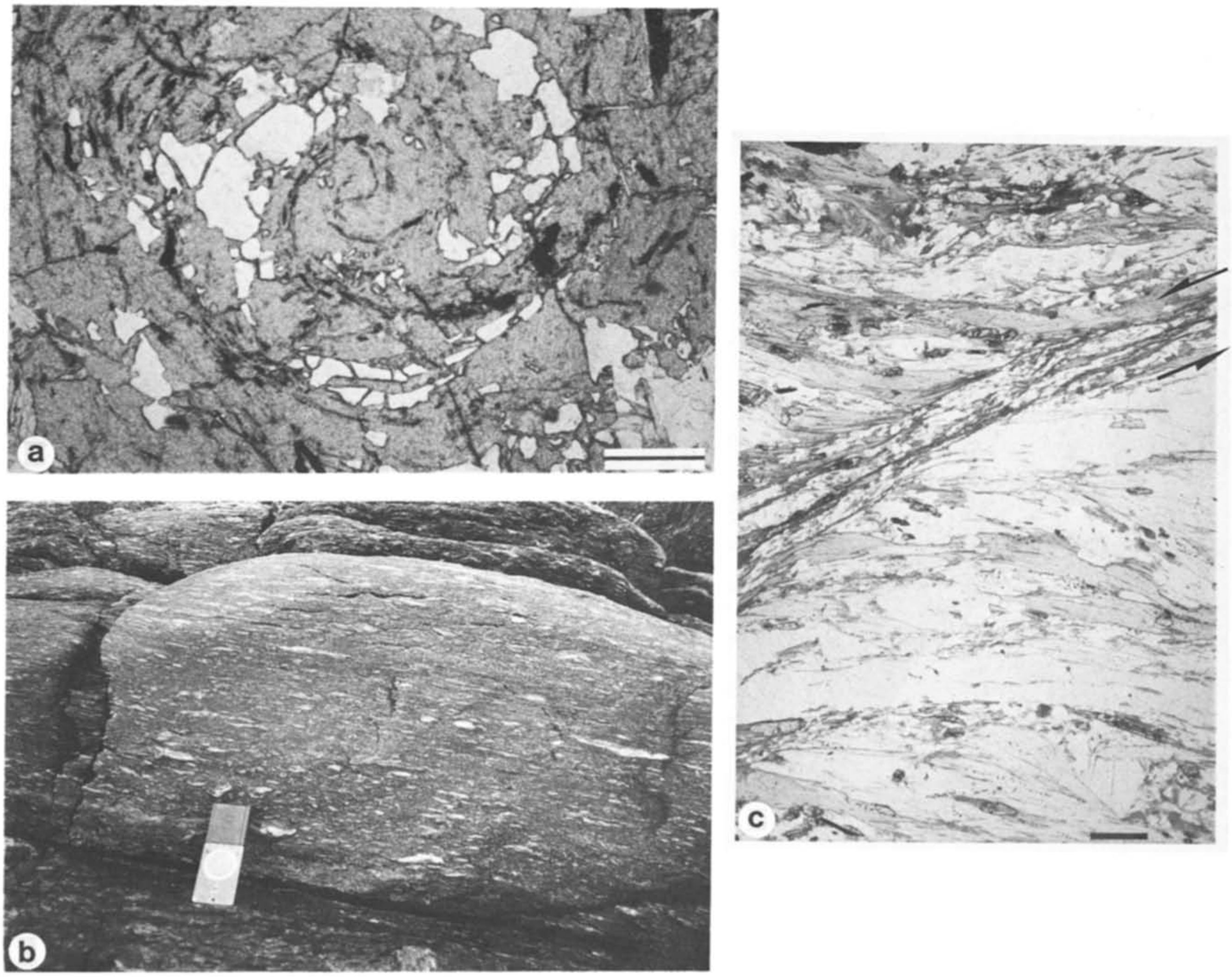


Fig. 7 (a) Centre of syn-D_3 garnet porphyroblast with inclusion trails of quartz and fine opaque material indicating rotation of over 360°. Scale bar is 0.5 mm. (b) Lenticular hornblende schist produced by intense D_3 shearing. (c) Shear band foliation cutting mica–schist. Scale bar is 0.1 mm.

Figure 3b shows the relationships between F_2, F_3, F_4 and F_5 folds at Balin Point on the east side of the island. The structural history of Dalziel (1984) contains fewer deformation phases than the sequence presented here. Dalziel's early phase (D_e) is probably equivalent to our D_2. His main phase (D_m) includes our D_3 and also folds we have mapped as D_4 structures. Similarly his late phase (D_l) includes some of our D_4 structures as well as the D_5 phase.

SYNTHESIS AND DISCUSSION OF THE DEFORMATION HISTORY

Early deformation

D_1 and D_2 deformation produced a layered schist with small-scale tight F_2 folds and a strong SSE-plunging lineation. This lineation comprises the L_2 rodding lineation of F_2 axes and boudins and the D_2 stretching lineation, X_2. L_2 is locally variable in trend but X_2 remains SSE-plunging although it may be obscured by later mineral lineations. Parallelism of tight fold axes and associated stretching lineation has been ascribed to rotation of axes from an initial orientation at a high angle to stretching during intense progressive shear (Bryant & Reed 1969, Sanderson 1973, Rhodes & Gayer 1977, Bell 1978, Quinquis *et al.* 1978, Berthé & Brun 1980, Cobbold & Quinquis 1980). Thus, the high D_2 strain, suggested by intrafolial folds and sheared out fold limbs and lithological contacts, may be due to intense shear sub-parallel to stretching, but the sense of shear cannot be deduced. The intense D_2 shearing produces a small-scale discontinuous rodding (Fig. 6a). No sheath folds were recognized but the ends of some discontinuous rods may be sheath fold noses. F_3 folds trend generally SSE but deviations of up to 90° occur and rare sheath folds, with long axes also trending SSE, are seen. Narrow shear zones, thin tectonic breccias and a SSE plunging mineral lineation were developed during intense D_3 deformation and syn-D_3 garnet porphyroblast rotation implies shear parallel to S_2 with a north-northwesterly vorticity. The deformation up to D_3 is thus consistent with a bulk shear parallel to S_2 and a NNW sense of overthrusting.

Late shearing

Flat-lying semi-pelitic mica–schists have an ubiquitous post-D_4 shear band foliation (White *et al.* 1980) dipping gently NW throughout the island (Figs. 7c and 8a). The foliation consists of 1 mm zones of shear with 1 cm spacing (Fig. 7c). Grain-size is reduced in the zones, producing a fine mosaic of quartz and feldspar and small mica flakes. Some biotite and chlorite grew along the shears but otherwise the deformation is post-metamorphic. The shear band foliation generally lies at 30–40° to the composite early foliation but sometimes curves into parallelism with it resulting in layer-parallel shearing. Slickensides are well developed and, together with offsets of the early foliation, they show that the shear

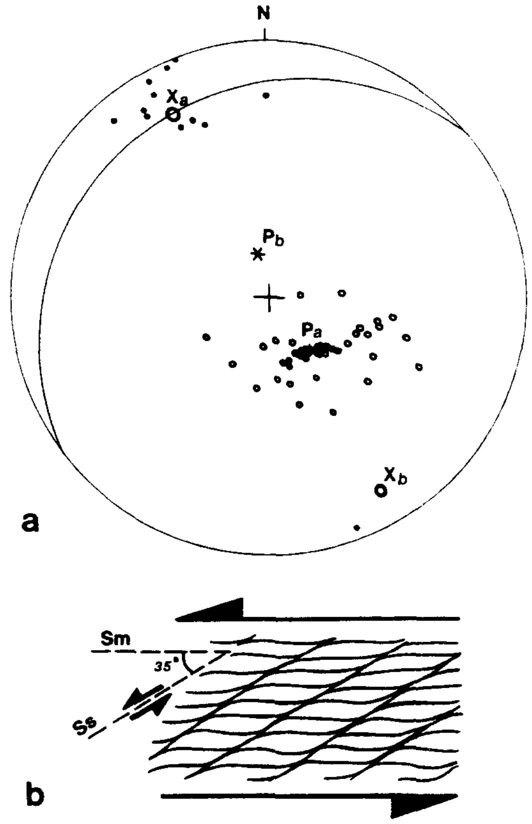


Fig. 8. (a) Poles to shear band foliation (○) and lineations parallel to shear direction (·) plotted on equal area stereogram. P_a, mean pole to shear band foliation with cyclographic trace. X_a, approximate shear direction on shear band foliation. Pole to bulk shear plane (P_b) and bulk shear direction (X_b) derived from P_a and X_a by rotation of 35°. (b) Relationship of shear band foliation (S_s) to main shear zone foliation (S_m; sub-parallel to bulk shear direction).

direction dipped at approximately 21° towards 334° (Fig. 8a) and the sense of displacement was that of low-angle, normal microfaults extending the foliation. In places the shear band foliation intensifies into shear zones 1–2 m wide. Sets of minor late extensional faults parallel to the shear band foliation are also developed. The shear bands are associated with grain-size reduction but not chemical differentiation or pressure solution. The deformation was thus probably volume constant and the foliation consists of micro-shear zones or S-bands (Cobbold 1977). An estimate of the shear strain involved can be made by measuring the angle between S_2 and the shear bands (α) and the angle S_2 makes with the shear bands after being displaced by several of them (α'). The shear strain (γ) is given by

$$\gamma = \cot \alpha' - \cot \alpha$$

and is approximately 1.7. The foliation can be penetrative through mica–schist up to 100 m thick and thus is responsible for large shear displacements.

White *et al.* (1980) described a shear band foliation at 35° to the main mylonite foliation in the Cap de Creus area of NE Spain. They suggested that it represents the final phase of ductile deformation in a shear zone and showed that it can be used to deduce the shear direction on the main mylonite foliation (Fig. 8b). Applying this geometrical relationship to the shear band foliation of

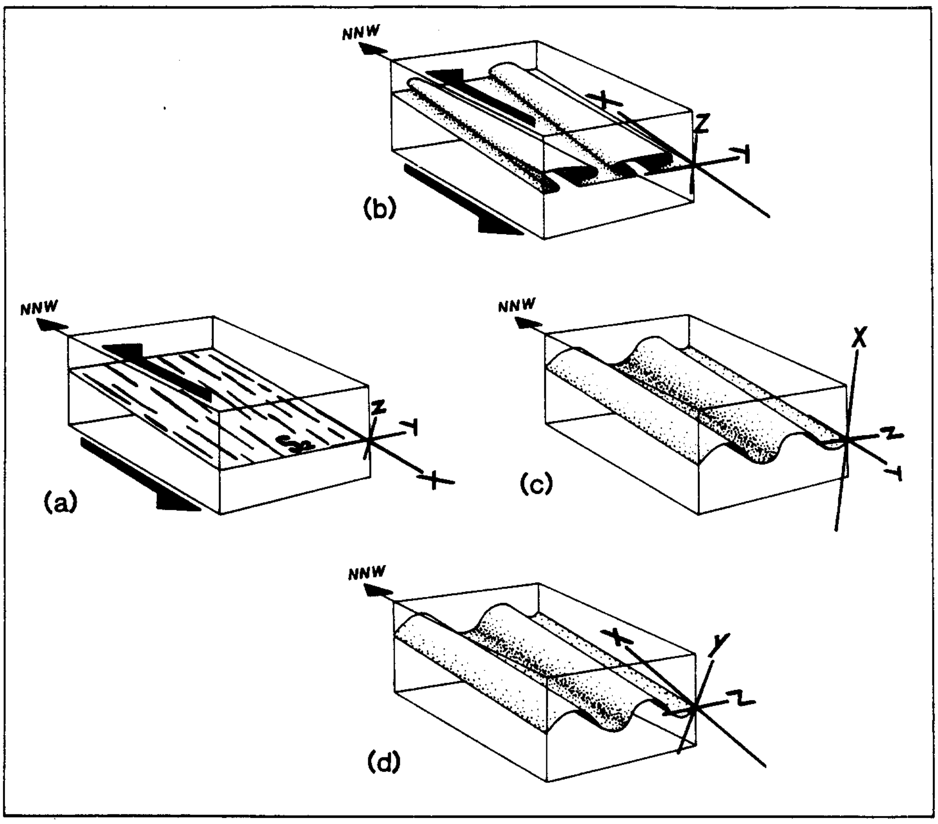


Fig. 9. (a) Finite strain at end of D_2 with rodding lineation sub-parallel to X. (b)–(d) Three possible strain states which could produce folds of S_2 with axes sub-parallel to X_2. See text for explanation.

Signy Island suggests a bulk thrusting direction inclined upwards at 13° towards 331° (Fig. 8a). Thus the bulk shear plane (P_b in Fig. 8a) is parallel to the composite early foliation, $S_{1/2}$, whilst the bulk shear direction (X_b in Fig. 8a) is slightly anticlockwise of the D_2 stretching direction (Fig. 5a).

It is likely that the shear band foliation, the minor shear zones, the layer-parallel shear and the D_5 folding are all expressions of the same late penetrative shearing. The type of structure which developed depended on the orientation of S_1 and on the lithology. In areas where S_1 is steep, F_5 folds are common: they are usually chevron-style in amphibolites and often have strongly sheared limbs in mica–schists. In areas of flat-lying S_1, layer-parallel shear, minor flat-lying shear zones and the oblique shear band foliation are developed, especially in mica–schists (e.g. in the Gourlay Peninsula).

The bulk shear direction implied by the shear band foliation is significantly oblique to F_4 and F_5 fold axes (Figs. 5b and 8a). The westerly directed component of simple shear tightens some F_4 folds and rotates their axial planes to a more gentle easterly dip. It also produces the common shearing on the limbs of F_5 folds.

Late folds parallel to stretching

Post D_2 fold phases all have axes sub-parallel to the early stretching and shearing direction [X sub-horizontal NNW–SSE (Fig. 9a)]. There are many possible strain states which could produce folds of this orientation in a flat-lying early composite schistosity and layering and these are summarized by the following four models.

(1) The strain axes could be inclined at moderate angles to the D_2 strain axes but the fold axes are controlled by a D_2 linear anisotropy (Fig. 9a).

(2) If simple shear continues in the same direction as during D_2, successive new fold axes might form at a high angle to stretching and be progressively rotated into parallelism with it (Fig. 9b) (Bryant & Reed 1969, Escher & Watterson 1974).

(3) Reorientation of the strain axes with X at a high angle to schistosity and Z at a low angle to it and trending WSW–ENE (Fig. 9c).

(4) X remains at a low angle to schistosity and either the deformation becomes constriction ($Y \approx Z$) or Z becomes oriented at a low angle to schistosity (Fig. 9d).

Cobbold & Watkinson (1981) and Watkinson & Cobbold (1981) have demonstrated that linear anisotropy in rocks can control the orientation of later folds (model 1). In the Signy Island rocks, however, the D_2 linear anisotropy, defined by the axes of small-scale tight folds and quartz rods, is unlikely to have been strong enough to control the orientation of major F_3 folds. Even minor F_4 folds refold the strongest D_2 rodding (Fig. 6a). It is likely, however, that the linear bending anisotropy resulting from F_3 folding has exerted some control on the attitude of F_4 and F_5 folds and that F_4 has affected F_5 orientation. The second model is discounted because late folds are often open, buckle folds and no F_4 or F_5 axes are seen at a high angle to the SSE trend. Although

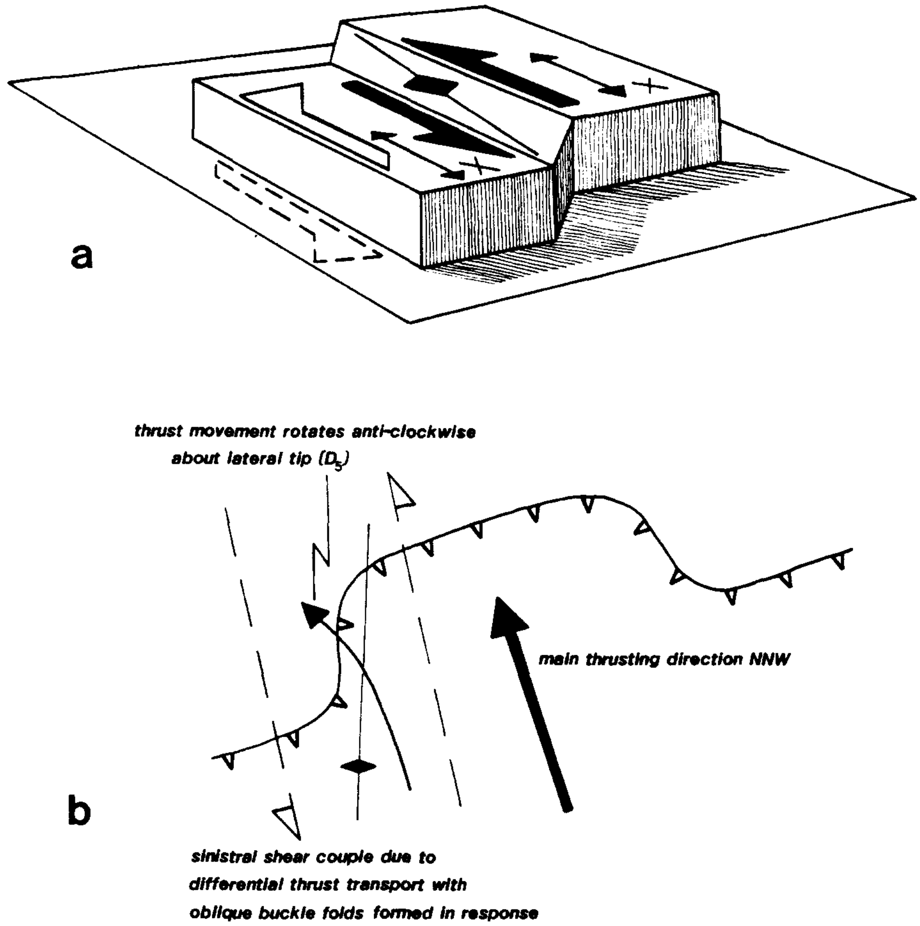


Fig. 10. (a) Development of fold axes sub-parallel to the thrusting direction (X) by differential thrust-transport (Coward & Kim 1981). (b) Differential thrust-transport model of Coward (1984) explains anti-clockwise rotation of D_5 shearing direction and formation of open F_3 and F_4 folds on Signy Island.

uncommon F_3 sheath folds suggest rotation of some axes, many F_3 folds are open and must have initiated with NNW trending axes. The D_3 stretching and shearing and the penetrative late shearing shows that NNW overthrusting was operative during D_3 and D_5; thus model 3 is discounted in favour of one where X remains at a low angle to foliation and trends NNW–SSE and the deformation is approximately simple shear. How the strain axes may have become modified during this progressive shear is discussed in the following section.

Modification of strain axes during thrusting to produce folds subparallel to the thrusting direction

Roberts (1974) described folds in the SW Highlands of Scotland with axes parallel to the early stretching lineation and restricted to zones of intense deformation. These folds (which he called $B' \perp B_1$ folds, following Voll (1960)) are polyclinal and open, with ill-defined axial planes and no axial-planar cleavage. He suggested that they formed by constriction in a plane normal to the stretching direction. Those F_4 folds on Signy Island unaffected by late shearing have open, often polyclinal profiles with no axial-planar cleavage and appear to have formed by buckling and parallel folding of S_2. They could have formed in the manner suggested by Roberts (1974) for his $B' \perp B_1$ folds.

Coward & Kim (1981) described late kink bands and chevron folds which plunge subparallel to the earlier stretching lineation associated with the Moine Thrust. They concluded that the late folds were too open in style to have been rotated into this direction by intense deformation and showed that their hinge-lines formed close to stretching where the rock carried a component of shear normal to the layering. This shear was ascribed to differential movement of the Moine Nappe (Fig. 10a). The development of structures in three dimensions during differential transport within thrust sheets has been modelled by Sanderson (1982). The differential transport is due to slip parallel to the shear direction and in a plane normal to the shear plane (analogous to a lateral ramp or wrench fault). As this lateral ramp is approached in a flat-lying thrust sheet the XY plane of the strain ellipsoid steepens and its strike rotates towards the transport direction. If a flat-lying shear zone foliation has formed during the early stages of thrusting, buckle folds of this foliation would be expected to form with axes close to X (Watkinson 1975) during differential transport at a later stage. Ridley (1982) described an example of the rotation and steepening of cleavage where localized wrench shears are superimposed on an overall thrust regime on Syros, Greece. On Signy Island the deformation is complex, there is a lack of strain markers and an absence of any systematic swing of

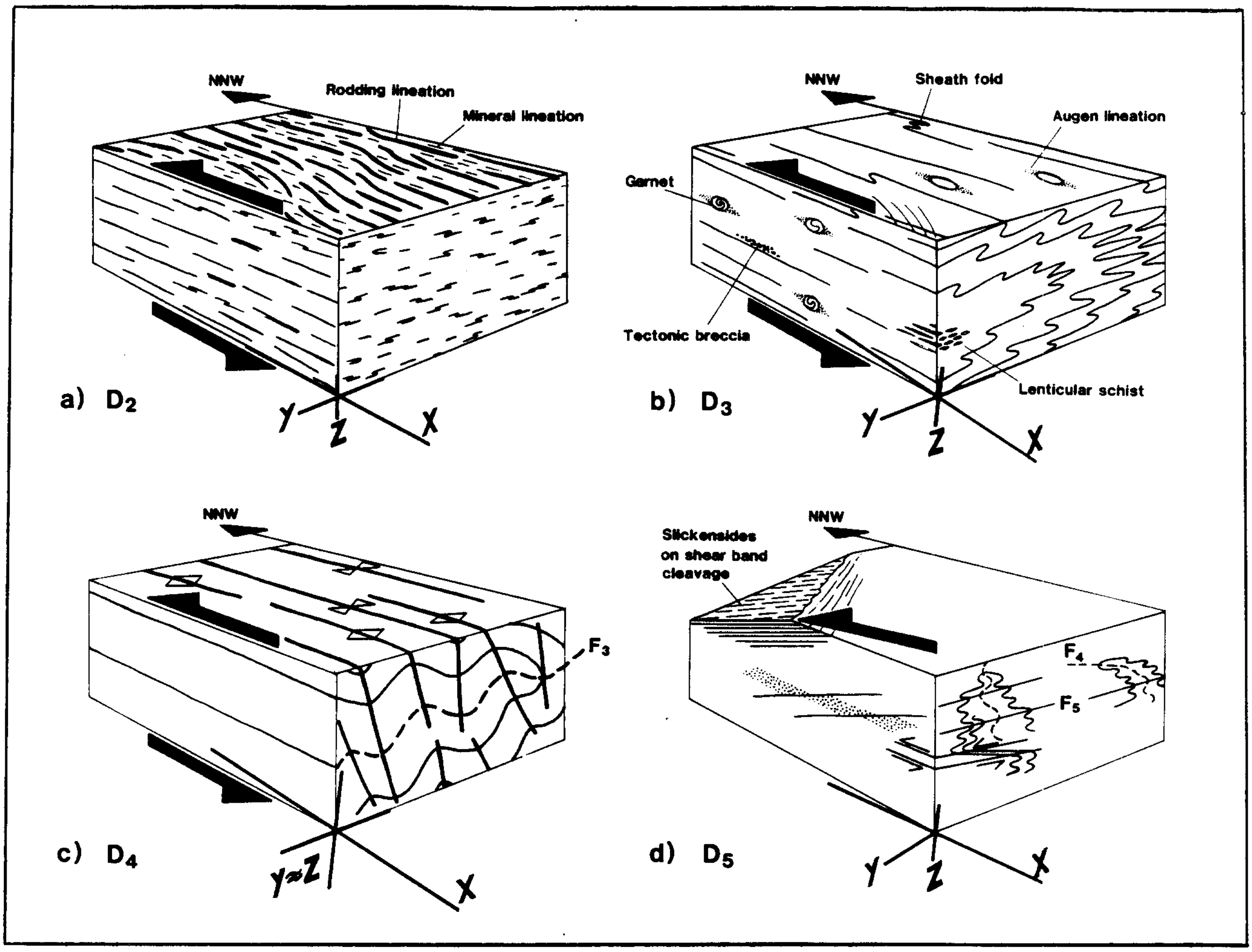


Fig. 11. Schematic models of the structures and inferred orientation of shearing direction and strain axes during D_2–D_5 deformations. (a) D_2: S_2 sub-parallel to shearing direction with well developed L_2 rodding lineation generally parallel to stretching (X). (b) D_3: Major F_3 fold. Sheath folds, tectonic breccia, lenticular schist, augen lineation and snowball garnet all suggest shearing in a NNW–SSE direction whilst garnet rotation indicates thrusting towards the NNW. (b) D_4: open folds develop with axes parallel to stretching. NNW thrusting modified by constrictional strain, differential thrust transport or lateral ramp. (d) D_5: flat-lying folds. Minor shear zones, shear band cleavage and extensional faults all indicate NW or NNW thrusting.

structures into a possible lateral ramp zone. This precludes quantitative analysis using the models of Sanderson (1982) but, in view of the consistent shearing direction during D_3 and D_5, differential thrust transport is a likely mechanism for the development of late open folds subparallel to X on the island. Alternatively, a major lateral or oblique ramp, which does not necessarily accommodate differential thrust transport, may exist in the vicinity of Signy Island. Folds formed in response to thrusts climbing the ramp will have axes parallel to the culmination walls. Syn-D_3 N–S trending faults (Fig. 6d) could be interpreted as minor lateral ramps.

Coward (1984) explained divergence of thrusting directions in the Moine Thrust zone at Assynt by pinning at a lateral tip resulting in a local rotation of the thrusting direction. We speculate that this model could explain the relations on Signy Island where the late (D_5) shearing is 15–20° anti-clockwise of the main early (D_2–D_3) thrusting direction and oblique folds are formed in response to the sinistral shear couple (Fig. 10b).

Jacobson (1983) described early isoclinal folds and sheath folds with axes parallel to stretching in schist below the Vincent Thrust, California. More open, later folds also have axes parallel to the early stretching and he concluded that this was evidence against any of the fold axes having rotated from an original trend at a high angle to stretching, although he demonstrated that folding took place during thrusting. We suggest that the structures noted by Jacobson (1983) are analogous to those on Signy Island and that the hinges of his early sheath folds have been rotated from a high angle to stretching (subparallel to the thrusting direction) into near-parallelism with this direction, whereas the late folds formed parallel to stretching due to modification of the strain axes within the zone of thrusting.

NNW OVERTHRUSTING THROUGHOUT THE DEFORMATION HISTORY ON SIGNY ISLAND

A model involving a constant NNW-directed shearing direction operating throughout the deformation on Signy Island to account for the consistent NNW–SSE alignment of fold axes, stretching and shearing is outlined below and in Fig. 11.

D_2 deformation produced small-scale tight folds, and intense progressive simple shear rotated F_2 axes into subparallelism with the S or SSE plunging stretching lineation (Fig. 11a). The sense of shear is unknown but is assumed to be overthrusting to the NNW as it was during the later phases. Thrusting to the NNW continued during D_3 deformation with large rotations of syntectonic garnet porphyroblasts, development of narrow intense shear zones and formation of sheath folds (Fig. 11b). Development of open F_3 folds with axes parallel to the stretching suggest a change in the Y and Z strain axes due to a constrictional deformation, lateral ramping or differential movement during thrusting. The variable trend of F_3 axes and the presence of sheath folds show that some F_3 axes were formed at a high angle to stretching and were subsequently modified by the progressive shear.

Formation of open folds with axes parallel to stretching continued during D_4 (Fig. 11c) and D_5 (Fig. 11d) whilst the pervasive late NNW directed shearing during and after D_5 produced the late shear band foliation, layer parallel shearing and minor shear zones.

DEFORMATION STYLE AT DEEP LEVELS IN AN ACCRETIONARY WEDGE

Thrusting related to subduction and accretion at high and intermediate levels in accretionary wedges has received much attention recently (for reviews see Karig & Sharman 1975 and references in Leggett 1982) and the accretionary process, particularly in younger fore-arc regions, is reasonably well understood. By contrast, there are few descriptions of thrust tectonics in medium or high-grade rocks at deep levels in subduction zones. Examples include the Ile de Groiz blueschists (Quinquis *et al.* 1978) and the deep-level thrust belt on Syros, Greece (Ridley 1982). These rocks are probably rare because they require uplift of 10 or 20 km to expose them and it may be difficult to distinguish them from rocks in ensialic ductile thrust belts or from rocks deformed during collision of an allochthonous terrane.

The Scotia metamorphic complex on the South Orkney Islands and the South Shetland Islands is considered to represent a subduction–accretion complex because of:

(1) its linear outcrop on the oceanic side of a magmatic arc,

(2) its lithologies and the occurrence of amphibolite derived from oceanic basalts (Storey & Meneilly 1985),

(3) the medium or high pressure metamorphic assemblage (Storey & Meneilly 1985) and presence of blueschists (Smellie & Clarkson 1975) and

(4) the intense shearing and tectonic interleaving (Dalziel 1984). Thus the structures described here from Signy Island may characterize the type of deformation in deep-level ductile thrust zones at the base of an accretionary wedge or in a zone of subduction.

IMPLICATIONS FOR PLATE MOVEMENTS

Marine geophysical data indicate that the Scotia Sea and the present configuration of the north and south Scotia ridges (Fig. 1) have evolved predominantly in the last 40–30 Ma (Barker & Griffiths 1972). Movement which may have taken place prior to 40 Ma is not recorded. Before this time there may have been continuous continental connection between South America and the Antarctic Peninsula (Hawkes 1962, Dalziel & Elliott 1971, Barker & Griffiths 1972, De Wit 1977). Fragmentation of this connection dispersed continental blocks eastwards to form the north and south Scotia Ridges.

Dalziel (1982, 1984) suggested that the South Orkney Islands microplate (containing Signy Island) can be rotated anti-clockwise to fit back into its original position occupying the present oceanic Powell Basin (Barker & Dalziel 1983). He based this reconstruction on matching bathymetry and compatibility of 'main phase' structural trend between the rocks of the restored South Orkney Islands microplate and their possible correlatives on southern Elephant Island in the South Shetland Islands (Fig. 12a). Stretching lineations are sub-parallel to hinge lines of the 'main-phase' folds on southern Elephant Island and the small islands to the south of it (Dalziel 1984). If these rocks are the same age as those in the South Orkney Islands and if the stretching direction on Elephant Island can be shown to be parallel to the shearing or thrusting direction then matching of structural trends would favour this reconstruction. However, the structural trends on the South Shetland Islands are variable and the age of the rocks is uncertain. Since the kinematic significance of the structural trends in the South Shetland Islands is also uncertain the present authors regard the matching of structural trends at this stage as unsatisfactory.

An alternative reconstruction is achieved by moving the South Orkney Islands block westwards with only minor anticlockwise rotation (Fig. 12b). This results in continuity of the South Orkney Islands continental margin with the arcuate trend of the Antarctic Peninsula continental margin. It also gives a better fit of the magnetic anomaly data (Harrington *et al.* 1972, Watters 1972), consistent with the pattern observable down the length of the Antarctic Peninsula (S. Garrett pers. comm.), than the reconstruction of Dalziel (1982 fig. 12.7). The alternative reconstruction (Fig. 12b) gives some overlap in present-day bathymetry but, due to fragmentation of the South Scotia Ridge, the bathymetry before breakup was probably not as it is today.

The present structural analysis has shown that the subduction complex rocks on Signy Island have suffered intense shearing throughout their deformation history in a consistent NNW–SSE direction during ductile thrusting towards the NNW. Throughout Coronation Island, the largest of the South Orkney Islands (Fig. 12), the dominant fold-axial trend and mineral lineation are parallel to the NNW–SSE trend of Signy Island (data from Dalziel (1984)). Thus the consistent stretching and

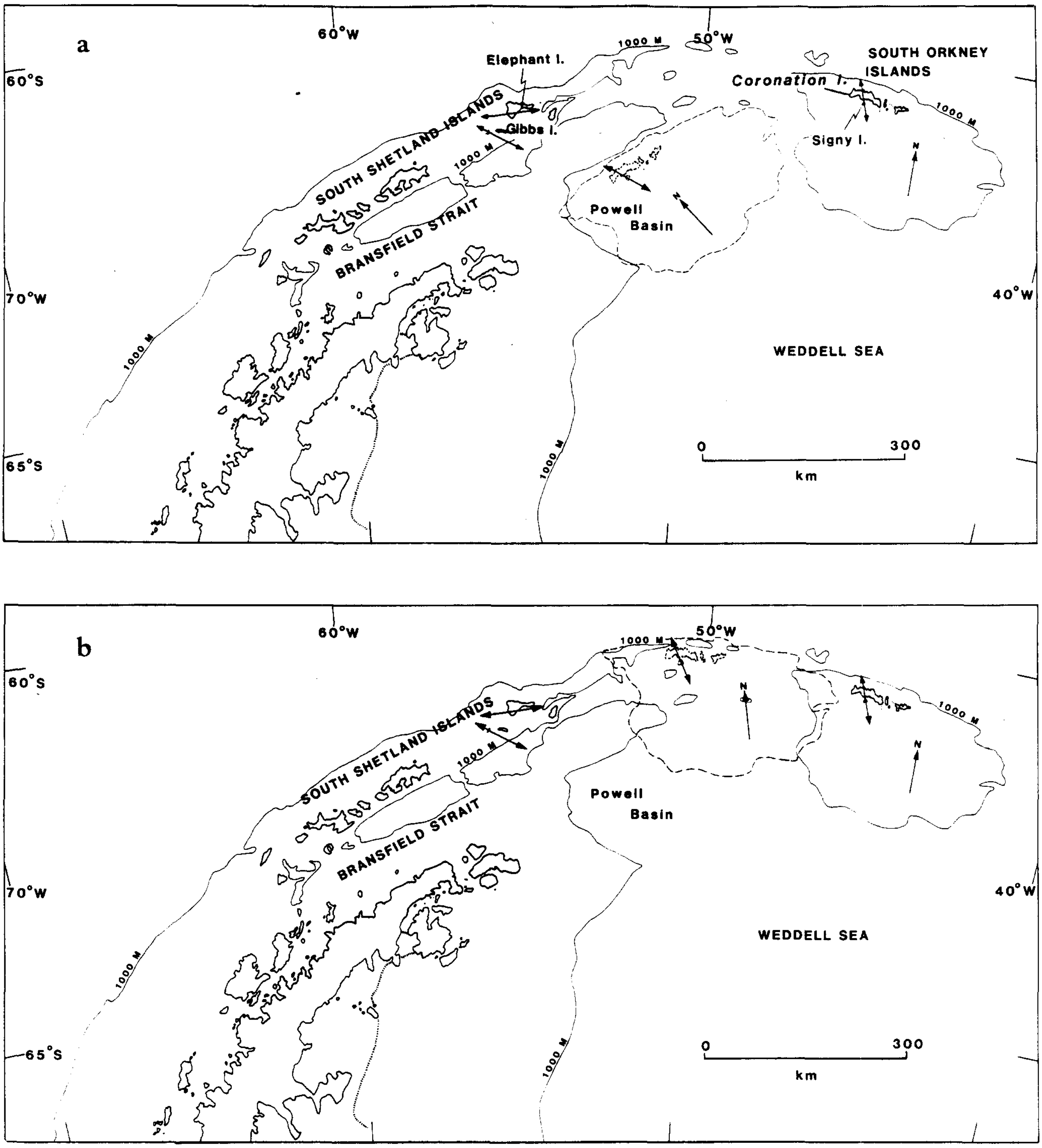


Fig. 12. (a) Reconstruction of original position of the South Orkney Islands block relative to the South Shetland Islands on the basis of matching bathymetry and structural trends (after Dalziel 1982, 1984). Late Cenozoic back-arc spreading in the Bransfield Strait (Ashcroft 1972) caused the South Shetland Islands to move away from the Antarctic Peninsula at right angles but this has been ignored in the reconstruction. (b) Restoration of the South Orkney Islands block westwards with only minor anticlockwise rotation resulting in continuity of the South Orkney Islands continental margin with the arcuate trend of the Antarctic Peninsula continental margin.

shearing direction on the small island of Signy probably indicates the thrusting direction in the South Orkney Islands block as a whole.

Stretching lineations have previously been used to infer plate motions in orogenic belts. For example, Quinquis *et al.* (1978) interpreted the stretching direction in subduction zone blueschists on the Ile de Groix as the direction of subduction. Shackleton & Ries (1984) showed that stretching lineations in the Himalayas coincide with the known relative motions of converging plates; they used this relationship to infer plate motions from the stretching lineations in the Variscan and Caledonian belts of Europe and in the late Proterozoic belts of north-east and east Africa. Thus, whilst acknowledging the possibility that shearing may be resolved into sequential movements transverse and parallel to the plate boundary (Shackleton & Ries 1984) the shearing direction in subduction complex rocks such as those on Signy Island could be used to infer the direction of subduction. The present work recognizes a consistent shearing direction and, by inference, subduction direction in the South Orkney Islands and this direction has kinematic significance and possible value in plate reconstructions. The restoration of the South Orkney Islands block presented here (Fig. 12b) suggests a subduction direction at a high angle to the arcuate trend of the Antarctic Peninsula continental margin. However, before assessing the alternative reconstructions, the significance and age of the structural trends in the South Shetland Islands must be more thoroughly investigated in order to identify the direction of tectonic transport.

Acknowledgements—We are grateful to the members of Signy base for their support during the field work. We thank our colleagues at the British Antarctic Survey, in particular S. W. Garrett, P. D. Marsh and J. W. Thomson for discussions and for critically reading the manuscript. I. W. D. Dalziel and P. W. G. Tanner are thanked for their constructive reviews of the paper.

REFERENCES

Ashcroft, W. A. 1972. Crustal structure of the South Shetland Islands and Bransfield Strait. *Scient. Rep. Br. Antarct. Surv.* No. 66.

Barker, P. F. 1982. The Cenozoic subduction history of the Pacific margin of the Antarctic Peninsula: ridge crest–trench interactions. *J. geol. Soc. Lond.* **139**, 787–801.

Barker, P. F. & Dalziel, I. W. D. 1983. Progress in geodynamics in the Scotia Arc region. In: *Geodynamics of the Eastern Pacific Region, Caribbean and Scotia Arcs* (edited by Cabre, R.). Geodynamics series, *Am. Geophys. Un.* **9**, 137–170.

Barker, P. F. & Dalziel, I. W. D. *et al.* 1976. The evolution of the southwestern Atlantic Ocean basin: Leg 36 data. In: *Init. Rep. Deep-sea Drilling Project*, U.S. Government Printing Office, Washington D.C. V **36**, 993–1014.

Barker, P. F. & Griffiths, D. H. 1972. The evolution of the Scotia Ridge and Scotia Sea. *Phil. Trans. R. Soc.* **A271**, 151–183.

Bell, T. H. 1978. Progressive deformation and reorientation of fold axes in a ductile mylonite zone: the Woodroffe thrust. *Tectonophysics* **44**, 285–320.

Berthé, D. & Brun, J. P. 1980. Evolution of folds during progressive shear in the South Armorican Shear Zone, France. *J. Struct. Geol.* **2**, 127–133.

Bryant, B. & Reed, J. C. 1969. Significance of lineation and minor folds near major thrust faults in the Southern Appalachians and the British and Norwegian Caledonides. *Geol. Mag.* **106**, 412–429.

Cobbold, P. R. 1977. Description and origin of banded deformation structures—I. Regional strain, local perturbations, and deformation bands. *Can. J. Earth Sci.* **14**, 1721–1731.

Cobbold, P. R. & Quinquis, H. 1980. Development of sheath folds in shear regimes. *J. Struct. Geol.* **2**, 119–126.

Cobbold, P. R. & Watkinson, A. J. 1981. Bending anisotropy: a mechanical constraint on the orientation of fold axes in an isotropic medium. *Tectonophysics* **72**, T1–T10.

Coward, M. P. 1984. The strain and textural history of thin-skinned tectonic zones: examples from the Assynt region of the Moine Thrust zone, NW Scotland. *J. Struct. Geol.* **6**, 89–99.

Coward, M. P. & Kim, J. H. 1981. Strain within thrust sheets. In: *Thrust and Nappe Tectonics* (edited by McClay, K. and Price, N. J.). *Spec. Publs Geol. Soc. Lond.*, 275–292.

Dalziel, I. W. D. 1982. The early (pre-Middle Jurassic) history of the Scotia arc region: a review and progress report. In: *Antarctic Geoscience* (Edited by Craddock, C.). University of Wisconsin Press, Madison, 111–126.

Dalziel, I. W. D. 1984. Tectonic evolution of a forearc terrane, southern Scotia Ridge, Antarctica. *Spec. Pap. geol. Soc. Am.* **200**.

Dalziel, I. W. D. & Elliott, D. H. 1971. Evolution of the Scotia arc. *Nature, Lond.* **233**, 246–252.

De Wit, M. J. 1977. The evolution of the Scotia arc as a key to the reconstruction of southwestern Gondwanaland. *Tectonophysics* **37**, 53–81.

De Wit, M. J., Dutch, S., Kligfield, R., Allen, R. & Stern, C. 1977. Deformation, serpentinization and emplacement of a dunite complex, Gibbs Island, South Shetland Islands: possible fracture zone tectonics. *J. Geol.* **85**, 745–762.

Dickinson, W. R. 1971. Plate tectonic models of geosynclines. *Earth Planet Sci. Lett.* **10**, 165–174.

Escher, A. & Watterson, J. 1974. Stretching fabrics, folds and crustal shortening. *Tectonophysics* **22**, 223–231.

Grikurov, G. E., Krylov, A. Y. & Silin, Y. I. 1967. Absolute age of some rocks from the Scotia arc and Alexander I Land (western Antarctica). *Dokl. Acad. Sci. U.S.S.R. Geol. Sci. Sect.* **172**, 19–22.

Harrington, P. K., Barker, P. F. & Griffiths, D. H. 1972. Crustal structure of the South Orkney Islands area from seismic refraction and magnetic measurements. In: *Antarctic Geology and Geophysics* (edited by Adie, R. J.). Universitetsforlaget, Oslo, 27–32.

Hawkes, D. D. 1962. The structure of the Scotia arc. *Geol. Mag.* **99**, 85–91.

Jacobson, C. E. 1983. Structural geology of the Pelona Schist and Vincent thrust, San Gabriel Mountains, California. *Bull. geol. Soc. Am.* **94**, 753–767.

Kanehira, K. 1967. Sanbagawa crystalline schists in the Iimori district, Kii Peninsula. *Jap. J. Geol. Geog.* **38**, 101–115.

Karig, D. E. & Kay, R. W. 1981. Fate of sediments on the descending plate at convergent margins. *Phil. Trans. R. Soc.* **A301**, 233–251.

Karig, D. E. & Sharman, G. F. III. 1975. Subduction and accretion in trenches. *Bull. geol. Soc. Am.* **86**, 377–389.

Leggett, J. K. (ed.) 1982. *Trench-Forearc Geology: Sedimentation and Tectonics on Modern and Ancient Active Plate Margins. Spec. Publs Geol. Soc. Lond.*

Matthews, D. H. & Maling, D. H. 1967. The geology of the South Orkney Islands—I. Signy Island. *F.I.D.S. Scientific Report* **25**.

Miller, J. A. 1960. Potassium–argon ages of some rocks from the South Atlantic. *Nature, Lond.* **187**, 1019–1020.

Quinquis, H., Audreu, C., Brun, J. P. & Cobbold, P. R. 1978. Intense progressive shear in Ile de Groix blueschists and compatibility with subduction or obduction. *Nature, Lond.* **273**, 43–45.

Rex, D. C. 1976. Geochronology in relation to the stratigraphy of the Antarctic Peninsula. *Bull. Br. Antarct. Surv.* **43**, 49–58.

Rhodes, S. & Gayer, R. A. 1977. Non-cylindrical folds, linear structures in the X direction and mylonite developed during translation of the Caledonian Kalak Nappe Complex of Finnmark. *Geol. Mag.* **114**, 329–341.

Ridley, J. 1982. Arcuate lineation trends in a deep level, ductile thrust belt, Syros, Greece. In: *Strain within Thrust Belts* (edited by Williams, G. D.). *Tectonophysics* **88**, 347–360.

Rivano, S. & Cortes, R. 1976. Note on the presence of the lawsonite-sodic amphibole association on Smith Island, South Shetland Islands, Antarctica. *Earth Planet. Sci. Lett.* **29**, 34–36.

Roberts, J. L. 1974. The structure of the Dalradian rocks in the S.W. Highlands of Scotland. *J. geol. Soc. Lond.* **130**, 93–124.

Sanderson, D. J. 1973. The development of fold axes oblique to the regional trend. *Tectonophysics* **16**, 55–70.

Sanderson, D. J. 1982. Models of strain variation in nappes and thrust sheets: a review. In: *Strain within Thrust Belts* (edited by Williams, G. D.). *Tectonophysics* **88**, 201–233.

Shackleton, R. M. & Ries, A. C. 1984. The relation between regionally consistent stretching lineations and plate motions. *J. Struct. Geol.* **6**, 111–117.

Smellie, J. L. 1981. A complete arc-trench system recognized in Gondwana sequences of the Antarctic Peninsula region. *Geol. Mag.* **118**, 139–159.

Smellie, J. L. & Clarkson, P. D. 1975. Evidence for pre-Jurassic subduction in western Antarctica. *Nature, Lond.* **258**, 701–702.

Storey, B. C. & Meneilly, A. W. 1985. Petrogenesis of metamorphic rocks within a subduction–accretion terrane, Signy Island, South Orkney Islands. *J. metamorphic Geol.* **3**, 21–42.

Suarez, M. 1976. Plate-tectonic model for southern Antarctic Peninsula and its relation to southern Andes. *Geology* **4**, 211–214.

Tanner, P. W. G., Pankhurst, R. J. & Hyden, G. 1982. Radiometric evidence for the age of the subduction complex in the South Orkney and South Shetland Islands, West Antarctica. *J. geol. Soc. Lond.* **139**, 683–690.

Thomson, J. W. 1968. The geology of the South Orkney Islands—II. The petrology of Signy Island. *Scient. Rep. Br. Antarct. Surv.* **62**.

Thomson, J. W. 1974. The geology of the South Orkney Islands—III. Coronation Island. *Scient. Rep. Br. Antarct. Surv.* **86**.

Voll, G. 1960. New work on petrofabrics. *Lpool. Manchr. Geol. J.* **2**, 503–597.

Watkinson, A. J. 1975. Multilayer folds initiated in bulk plane strain, with the axis of no change perpendicular to the layering. *Tectonophysics* **28**, T7–T11.

Watkinson, A. J. & Cobbold, P. R. 1981. Axial directions of folds in rocks with linear/planar fabrics. *J. Struct. Geol.* **3**, 211–217.

Watters, D. G. 1972. Geophysical investigation of a section of the south Scotia Ridge. In: *Antarctic Geology and Geophysics* (edited by Adie, R. J.). Universitetsforlaget, Oslo, 33–38.

White, S. H., Burrows, S. E., Carreras, J., Shaw, N..D. & Humphreys, F. J. 1980. On mylonites in ductile shear zones. *J. Struct. Geol.* **2**, 175–187.

Wood, B. L. 1978. The Otago Schist mega-culmination: its possible origins and tectonics significance in the Rangitata Orogen of New Zealand. *Tectonophysics* **47**, 339–368.