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Tectonophysics 319 (2000) 93–110

TECTONOPHYSICS

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Deformational evolution of a Cretaceous subduction complex: Elephant Island, South Shetland Islands, Antarctica

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Received 25 May 1998; accepted for publication 7 January 2000

Abstract

New structural data from Elephant Island and adjacent islands are presented with the objective to improve the understanding of subduction kinematics in the area northeast of the Antarctic Peninsula. On the island, a first deformation phase, D_1, produced a strong SL fabric with steep stretching and mineral lineations, partly defined by relatively high pressure minerals, such as crossite and glaucophane. D_1 is interpreted to record southward subduction along an E–W trench with respect to the present position of the island. A second phase, D_2, led to intense folding with steep E–W-trending axial surfaces. The local presence of sinistral C′-type shear bands related to this phase and the oblique inclination of the L_2 stretching lineations are the main arguments to interpret this phase as representing oblique sinistral transpressive shear along steep, approximately E–W-trending shear zones, with the northern (Pacific) block going down with respect to the southern (Antarctic Peninsula) block. The sinistral strike-slip component may represent a trench-linked strike-slip movement as a consequence of oblique subduction. Lithostatic pressure decreased and temperature increased to peak values during D_2, interpreted to represent the collision of thickened oceanic crust with the active continental margin. The last deformation phase, D_3, is characterised by post-metamorphic kink bands, partially forming conjugate sets consistent with E–W shortening and N–S extension. The rock units that underlie the island probably rotated during D_3, in Cenozoic times, together with the trench, from an NE–SW to the present ENE–WSW position, during the progressive opening of the Scotia Sea. The similarity between the strain orientation of D_3 and that of the sinistral NE–SW Shackleton Fracture Zone is consistent with this interpretation. © 2000 Elsevier Science B.V. All rights reserved.

Keywords: Accretionary wedge; Oblique subduction; Polyphase deformation; South Shetland Islands; Subduction complex

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PII: S0040-1951(00)00021-4

1. Introduction

South America and the Antarctic Peninsula are connected by a large arcuate structure, the Scotia arc (Fig. 1). This arc encloses a complex system of microplates which formed during the Cenozoic as a consequence of the break-up of a continuous active margin that connected South America and the Antarctic Peninsula (Barker et al., 1991). Development of the Scotia arc is relatively well understood and can be reconstructed from the present geometry of the arc and the structure of the ocean floor of the Scotia Sea (Barker et al., 1991; Cunningham et al., 1995). The Mesozoic to Cenozoic kinematic history of the active margin mentioned above is less well understood, since it must be reconstructed from deformed and metamorphosed rock sequences only preserved on a few scattered islands along the Scotia arc. Along the South Scotia Ridge these islands are Smith Island, the Elephant Island group and the western South Orkney Islands (Fig. 1). The metamorphic rocks on these islands have been collectively named the Scotia metamorphic complex (Tanner et al., 1982; Dalziel, 1984), interpreted as a Mesozoic–Cenozoic subduction complex (Smellie and Clarkson, 1975; Dalziel, 1984; Grunow et al., 1992). The structure of this complex, or parts of it, have been described by Dalziel (1984), Meneilly and Storey (1986), Trouw and coworkers (Trouw, 1988; Trouw et al., 1997) and Grunow et al. (1992). The complex is now generally considered as composed of three different parts, with different metamorphic ages (e.g. Trouw et al., 1998a): (1) western South Orkney Islands with ages in the range 200–180 Ma; (2) Elephant Island group with ages between 90 and 110 Ma; (3) Smith Island with an age around 50 Ma.

The objective of this paper is to present a new tectonic interpretation of the metamorphic complex that crops out on Elephant Island (61°15′S, 55°00′W), situated in the South Shetland Islands (Fig. 1). The rock successions of this island play a key role in the reconstruction of the Mesozoic regional tectonic history because they suffered their main deformation and metamorphism in the Cretaceous (Tanner et al., 1982) and because several metamorphic zones are exposed, revealing an oblique section through most of the subduction complex (Fig. 2; Trouw et al., 1998a).

Although Elephant Island is situated close to the Shackleton Fracture Zone and the South Scotia Ridge (Fig. 1), both active transform plate boundaries (Klepeis and Lawver, 1996; Kim et al., 1997), the rocks that crop out at the island seem to have survived post-metamorphic deformation essentially as a coherent block. This is evident from the occurrence of only a few late brittle structures that might be associated with these fracture zones and from the metamorphic pattern, showing gradual transitions between zones and facies, without major disruptions (Fig. 2a).

This study is based on data collected in the austral summers of 1991–1992 and 1995–1996. Elephant Island, named after the southern elephant seal, is 90% ice covered and consists of a plateau with some ridges and isolated mountains up to about 1000 m high. The island has been uplifted relatively recently by about 100 m, creating steep coastal cliffs with hanging glaciers. These cliffs provide excellent outcrop along most of the coast, and allow detailed structural analysis. Some scattered nunataks in the interior, of difficult access, proved important to complete the structural and metamorphic analyses.

2. Lithology

The metamorphic succession that crops out at Elephant Island comprises grey, green and blue phyllites and schists, with local intercalations of thin beds of metachert, calc-silicate rocks and marble, and thin to very thick layers of amphibolite and fine volcanic metaconglomerate (e.g. Marsh and Thompson, 1985; Trouw et al., 1991). Whole-rock chemical analyses of mafic samples indicate them to be derived from an ocean floor environment (Valeriano et al., 1997). Most of the phyllites and rocks of chemical origin are interpreted as (hemi)pelagic sediments also from the ocean floor (Dalziel, 1984; Marsh and Thompson, 1985; Trouw et al., 1991; Grunow et al., 1992). Part of the phyllites and schists may be arc-derived turbidites (Heilbron et al., 1995). The fine volcanic conglomerates and associated metasandstones may

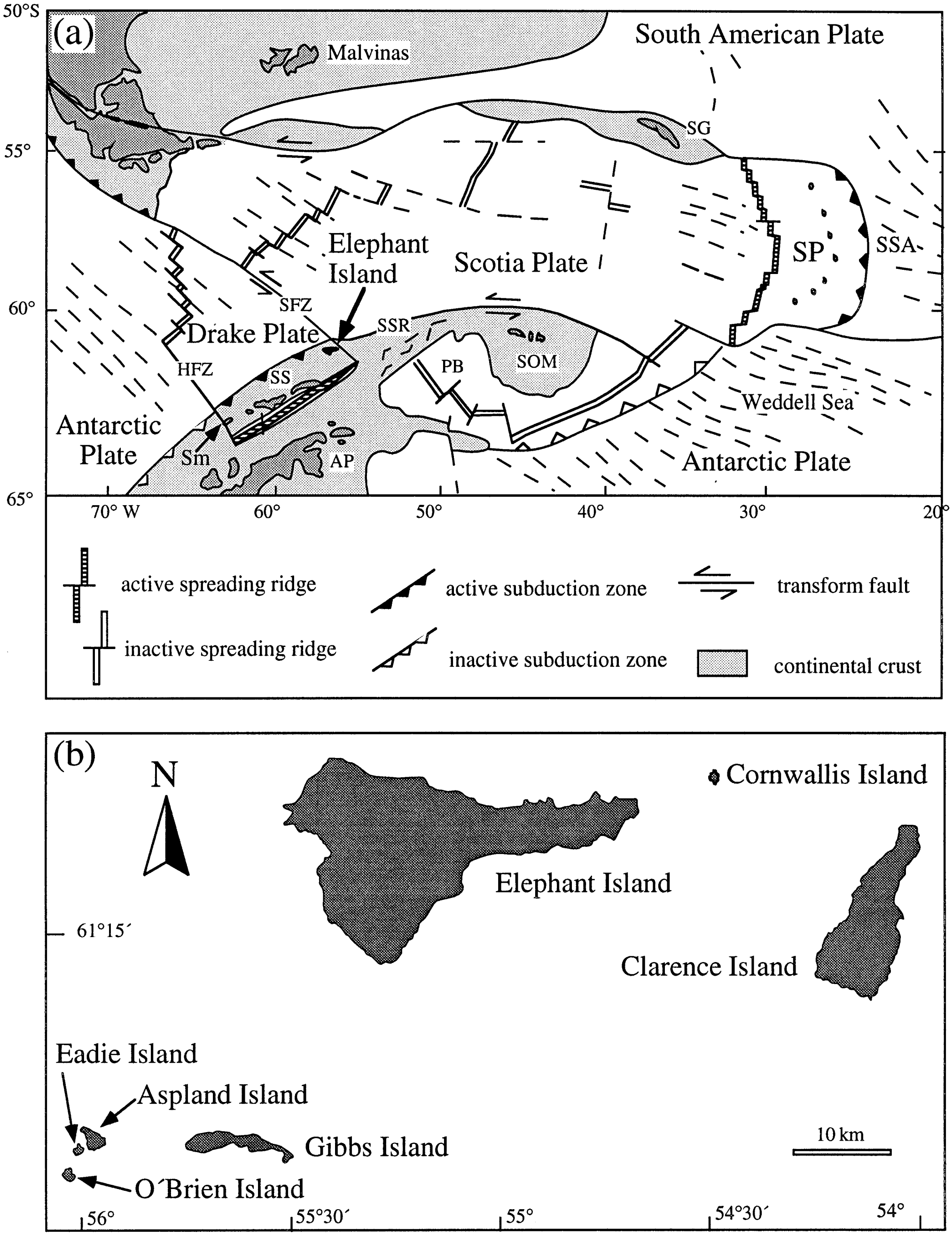
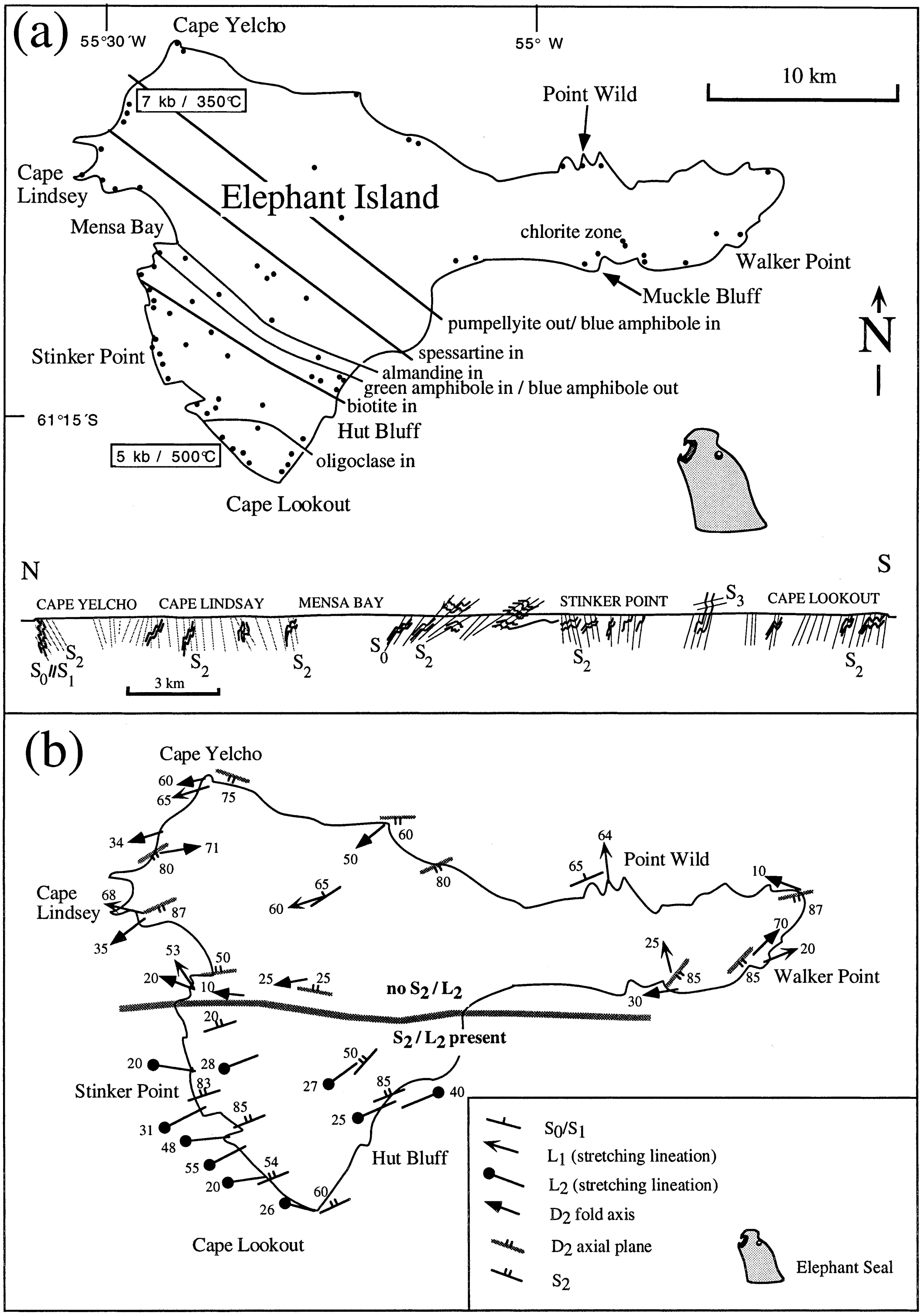


Fig. 1. (a) Geotectonic map of the Scotia arc with location of Elephant Island. AP: Antarctic Peninsula; HFZ: Hero Fracture Zone; PB: Powell Basin; SFZ: Shackleton Fracture Zone; SG: South Georgia; SM: Smith Island; SOM: South Orkney microplate; SP: Sandwich Plate; SS: South Shetland Islands; SSA: South Sandwich arc; SSR: South Scotia Ridge Dashed lines are fracture zones. (b) Location of islands of the Elephant Island group.


(a)
55°30´W
55° W
61°15´S
Cape Yelcho
7 kb / 350°C
Point Wild
10 km
Cape Lindsey
Elephant Island
Mensa Bay
chlorite zone
Walker Point
Muckle Bluff
N
pumpellyite out/ blue amphibole in
Stinker Point
spessartine in
almandine in
green amphibole in / blue amphibole out
biotite in
Hut Bluff
5 kb / 500°C
oligoclase in
Cape Lookout
N
S
CAPE YELCHO
CAPE LINDSAY
MENSA BAY
STINKER POINT
CAPE LOOKOUT
S_3
$S_0//S_1$
S_2
S_0
3 km
(b)
Cape Yelcho
Cape Lindsey
Point Wild
Walker Point
no S_2/L_2
S_2/L_2 present
Stinker Point
Hut Bluff
Cape Lookout
S_0/S_1
L_1 (stretching lineation)
L_2 (stretching lineation)
D_2 fold axis
D_2 axial plane
S_2
Elephant Seal

be derived from the erosion of a volcanic seamount.

3. Metamorphism

The metamorphism, described in detail by Trouw et al. (1998a,b), shows similarities with Sanbagawa-type metamorphism from Japan (Dalziel, 1984; Trouw et al., 1991; Grunow et al., 1992). The Elephant Island metamorphic succession shows a gradual transition from rocks belonging to the pumpellyite–actinolite facies in the northeast (Fig. 2a), through crossite–epidote–blueschist facies in the central and northwestern part, greenschist facies in the southwestern part and finally to amphibolite facies around Cape Lookout in the extreme south (Trouw et al., 1998a,b). These transitions are marked by six isograds: (1) pumpellyite out-blue amphibole in; (2) spessartine in; (3) almandine in; (4) green amphibole in-blue amphibole out; (5) biotite in; (6) oligoclase in. Most metamorphic minerals analysed by K–Ar, Rb–Sr and Ar–Ar methods yielded Cretaceous ages of 90–110 Ma (Tanner et al., 1982; Trouw et al., 1990; Hervé et al., 1991; Grunow et al., 1992). Comparison of core- and rim-compositions of metamorphic minerals, mainly garnet, amphibole and biotite, in different samples, and microtectonic analysis led to the reconstruction of clockwise *P–T–t* paths (Trouw et al., 1998a,b). Although these *P–T–t* paths show significant differences from north to south, their relation with the sequence of deformation phases is similar: increasing *P* and *T* during the first deformation phase (D_1); decreasing *P* and increasing *T* during the second deformation phase (D_2); and, finally, low *P* and *T* during the third deformation phase (D_3). D_1 and D_2 are both roughly contemporaneous with the Cretaceous metamorphism. The early, relatively high-pressure low-temperature stage of the metamorphism (~7 kbar, 350°C; Trouw et al., 1998a) related to D_1 is attributed to a subduction setting because of the corresponding low thermal gradient of about 17°/km. The higher temperature and somewhat lower pressure conditions, during D_2 (~5 kbar, 500°C), recorded in the southern part of the island, are thought to result from collision of an oceanic plateau (docking of a minor terrane) with the active continental margin (Trouw et al., 1998a).

4. Deformation structures

Macro- and mesoscopic structures on Elephant Island were described by Dalziel (1984), Trouw (1988) and Grunow et al. (1992). The structures of the Scotia metamorphic complex were subdivided into three groups (Dalziel, 1984; Grunow et al., 1992), attributed to three deformational phases: early phase (D_E), main phase (D_M) and late phase (D_L). Trouw (1988) used symbols D_1, D_2 and D_3 for these phases and this nomenclature is maintained here.

4.1. D_1 structures

D_1 structures, well preserved in the lower-grade metamorphic zones, are a slaty cleavage, S_1, a stretching and/or mineral lineation, L_1, developed on S_1, and tight to isoclinal folds of bedding (S_0). The strongly elongated prolate shape of detrital clasts in the volcanic metaconglomerate (strain ratios between principal axes X/Z up to 9:1; Trouw, 1988), associated with micro-boudinage and stretched minerals demonstrate that L_1 is a stretching lineation developed at high strain. D_1 folds are common, especially in thinly laminated metachert layers, and often show a sheath-like appearance (Trouw et al., 1991). Advanced transposition produced a situation in which S_1 is usually

Fig. 2. (a) Metamorphic map of Elephant Island with location of isograds and simplified structural section along the west coast. Black dots represent studied outcrops with analysed thin sections. Isograds refer essentially to peak temperature conditions, which were attained during D_2; see Trouw et al. (1998a) for further details. (b) Simplified structural map of Elephant Island. The E–W line marked on the map separates the southern part of the island, where S_2 and L_2 are usually present, from the northern part, where these structural elements are not developed.

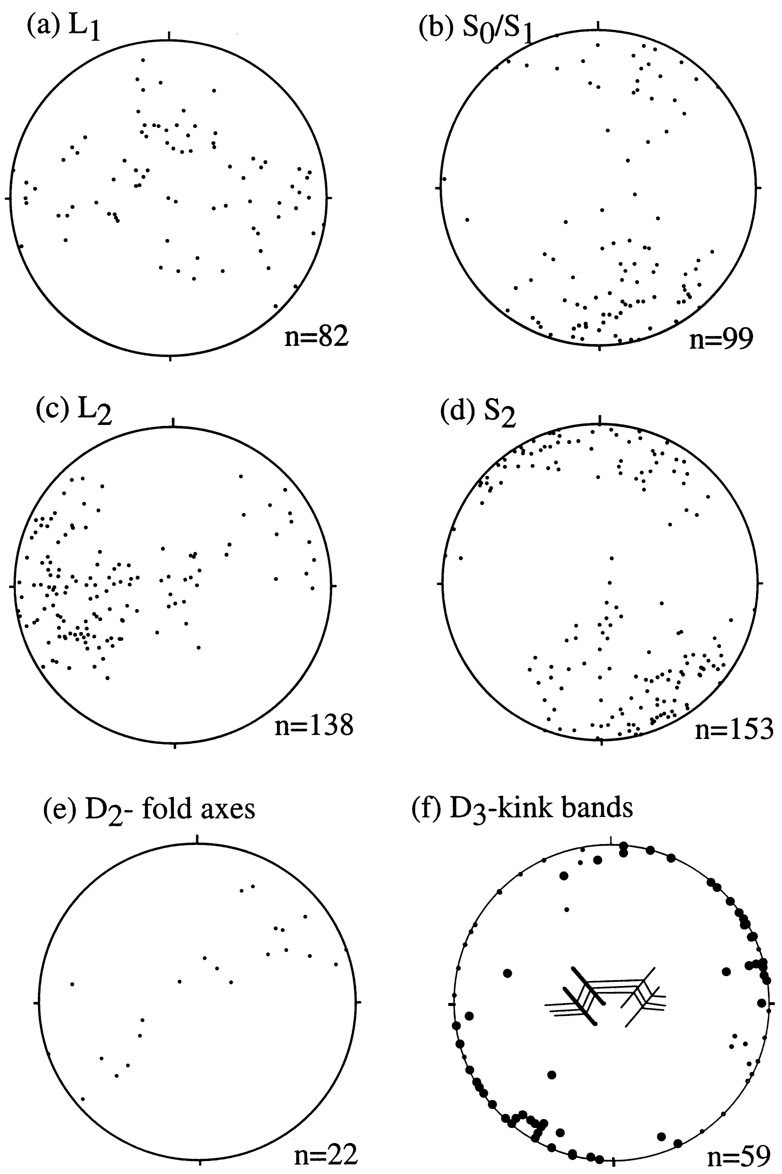


Fig. 3. Stereograms of structural measurements at Elephant Island. (a)–(e) Structural elements as indicated. S_0/S_1 refers to measurements of bedding (S_0) subparallel to S_1 cleavage. (f) Poles to two sets of conjugate kink bands with steep axes. Small and large dots indicate opposite kink geometries, as illustrated in the centre of the diagram. Equal area projection, lower hemisphere.

subparallel to S_0; a clear angle between the two surfaces could only be measured locally. The attitude of S_1 is generally steep, roughly with E–W strike and predominantly north dipping (Figs. 2b and 3b); L_1 has a variable orientation with both high and low rakes (Figs. 2b and 3a). The few D_1 fold axes that could be measured are subparallel to L_1.

Over most of the island the D_1 structures are modified by strong D_2-overprinting. However, around Point Wild, on the north coast (Fig. 2) the D_1 structures are well preserved in metaconglomerates and metasandstones. S_0 (bedding) is dipping to the NNW (347/65), subparallel to S_1, in a normal (top upwards) position (Fig. 2b). L_1, well defined at this site by stretched grains and pebbles, is north-plunging (002/64) with a high rake.

4.2. D_2 structures

Structures attributed to D_2 predominate over D_1 structures in the higher-grade metamorphic zones (Fig. 2). D_2 folds, deforming S_0 and S_1, are open to tight and approximately cylindrical. Interference patterns of D_1 folds refolded by D_2 are locally developed (Dalziel, 1984; Trouw, 1988). At some places, especially in quartz veins, a well developed L_2 stretching lineation defined by elongate quartz crystals or aggregates, and elongate micas is present, approximately parallel to D_2 fold axes.

D_2 structures show a clear gradient in style across the succession from north to south, expressed by an increasing tightness of D_2 folds and crenulations, and by the progressive development of S_2 through transposition of S_1 [compare fig. 4.17 in Passchier and Trouw (1996)], and increasing intensity of L_2 development. At Point Wild, in the north, hardly any D_2 deformation is present and at most other outcrops in the northern part of the island D_2 folds are open, without axial planar cleavage. By contrast, all outcrops in the southern part show a well-developed penetrative schistosity (S_2) and usually tight D_2 folds (Fig. 2b), commonly with remnants of tight crenulations of

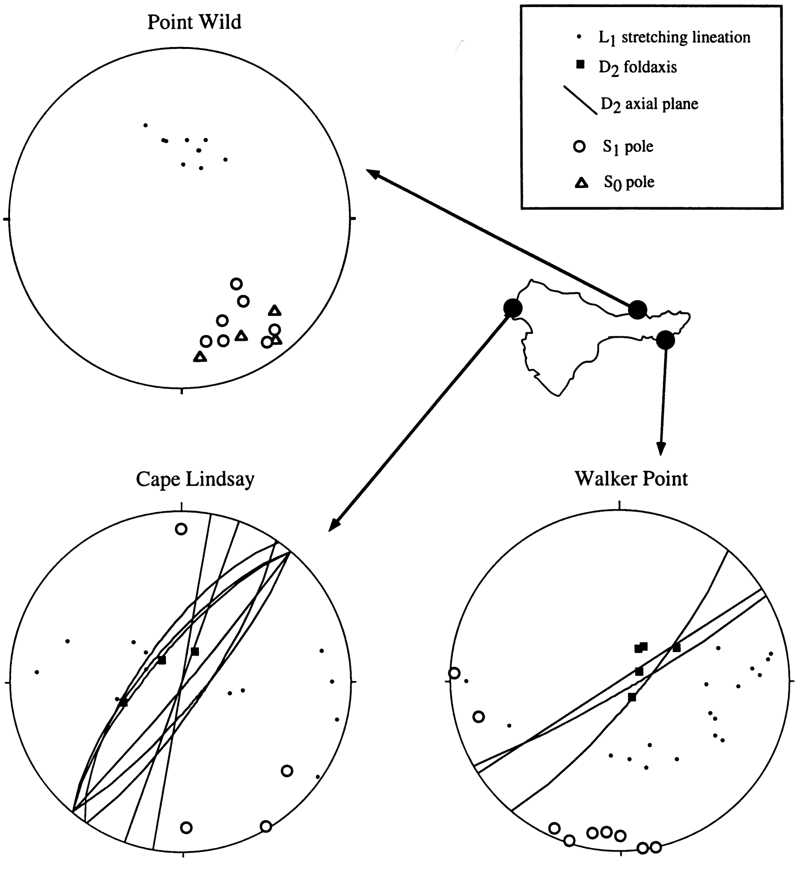


Fig. 4. Synoptical stereograms of several structural elements at three sites of northern Elephant Island, showing progressive rotation of L_1 stretching lineations from a N-plunging position to an EW-girdle.

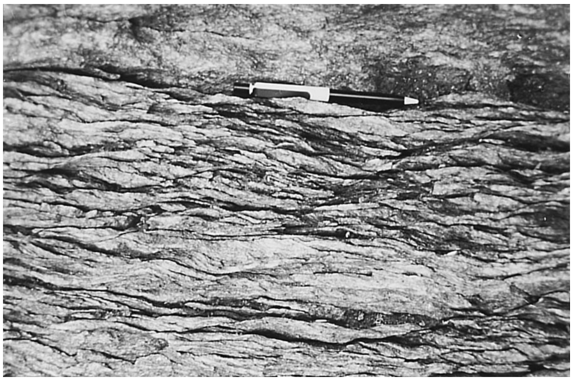


Fig. 5. C′-type shear band cleavage related to S_2 cleavage, indicative of sinistral shear movement.

S_1. This trend is interpreted as the effect of a southward increase in D_2 strain-intensity accompanied by an increase in metamorphic temperature (Trouw et al., 1998a,b), evident from the pattern of metamorphic isograds (Fig. 2a).

Around Stinker Point (Fig. 2) L_1 is often folded in D_2 folds, resulting in the presence of two clearly distinct lineations: a folded L_1 and a straight L_2, parallel to D_2 fold hinges. Further south, only one stretching lineation exists parallel to the hinges of D_2 folds. The generation of this lineation is interpreted to be a combination of rotated L_1 and newly formed L_2 lineation elements.

Fig. 3 shows that the distribution of L_1 attitudes forms a girdle that trends slightly clockwise from the girdle of L_2 orientations (Figs. 3 and 4). A tentative explanation for this fact is given below in Section 7.

D_2 axial planes and, in the south, S_2, strike roughly WSW–ENE (Figs. 2b and 3d) with steep dips, mostly to the NNW, in the northern and southern parts. Gently north-dipping S_2 appears in the central part of the island (Fig. 2a). L_2 and D_2 fold hinges are WSW–ENE trending, with a dominance of gently to steeply WSW plunging orientations (Fig. 3c). The mean value of the L_2 plunge is about 30° to the WSW. At Hut Bluff and Walker Point (Fig. 4), ENE-plunging lineations and fold axes predominate.

At several places where D_2 deformation is strong (e.g. Stinker Point), a C′-type shear band cleavage (Passchier and Trouw, 1996) is developed at a small angle to S_2 planes (Figs. 5 and 6). The intersection lineation of this cleavage with S_2 is highly oblique or orthogonal to L_2. The presence and orientation of the C′-shear band cleavage indicates that D_2 was a phase of non-coaxial flow, probably with its main displacement direction parallel to L_2. Shear sense indicated by the shear bands is invariably sinistral looking down along the steeply inclined intersection lineation. The inclined position of L_2 implies that the D_2 sinistral shear movements had a significant dip-slip component with relative uplift of the southern block

4.3. D_3 structures

Kink bands, often in conjugate sets of centimetre- to metre-scale, deforming both S_1 and S_2, have been observed locally throughout the whole succession They formed obviously much later than D_2, after the rocks had been uplifted to non-metamorphic or very low-grade near-surface conditions. The kink bands and their axes are steep in the southern part of the island (Fig. 7), where the S_2 foliation is also steep and well-developed. In other parts of the island the kink bands have a more irregular orientation. The preferred orientation pattern (Fig. 3f) shows that many kinks are consistent with E–W shortening and N–S extension with respect to the present orientation of the island. Locally, other conjugate sets of inclined kink bands with subhorizontal axes indicate vertical shortening.

5. Relation between deformation phases and metamorphism

The L_1 mineral lineation is defined by acicular minerals, including relatively high-pressure amphiboles such as glaucophane, crossite and barroisite (Trouw et al., 1998a,b). The high degree of alignment with L_1 indicates that these crystals must have grown during D_1.

Garnets with spiral-shaped inclusion patterns, continuous with S_1 (Fig. 8), were interpreted as syntectonic with respect to D_1. Other crystals have cores with spiral-shaped inclusion patterns and outbowing S_2 schistosity included in the rims that,

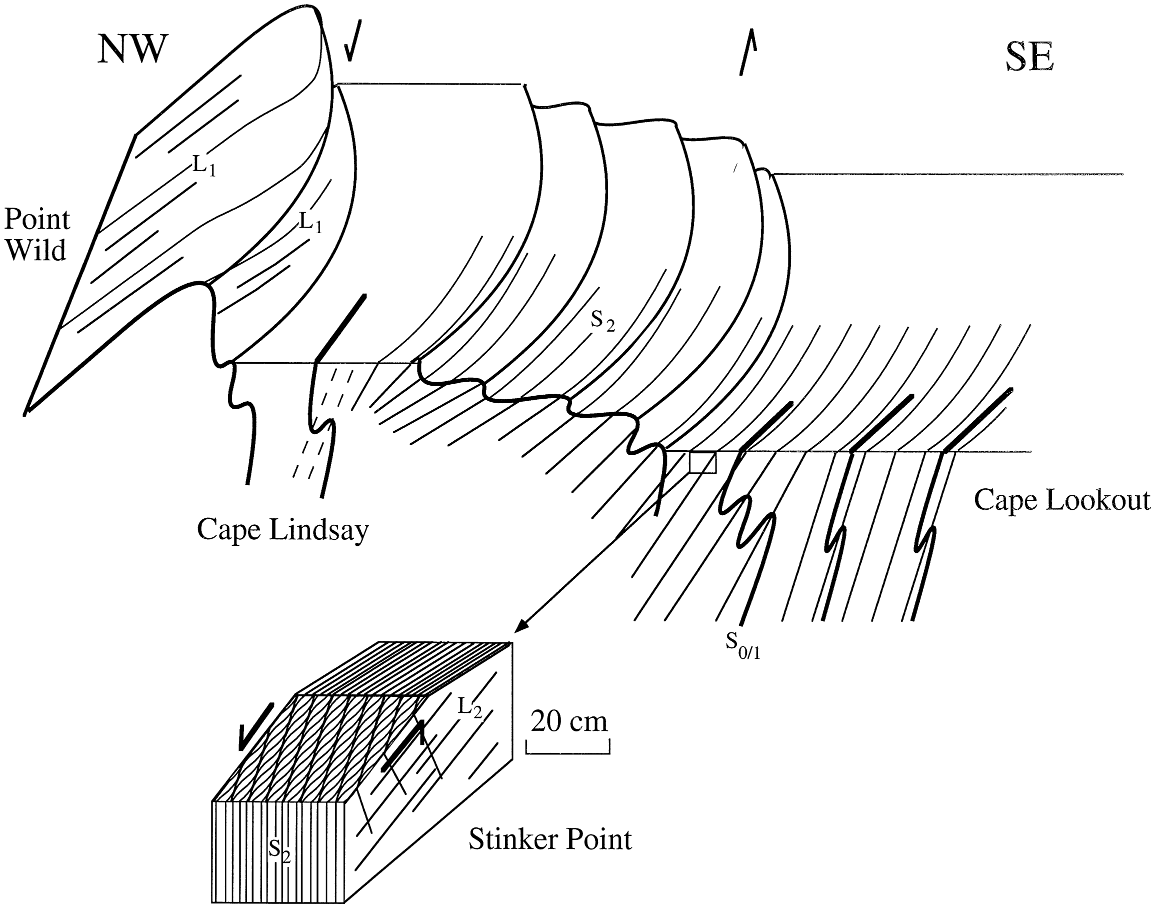


Fig. 6. Simplified structural scheme of Elephant Island showing mainly D_2 structures with increasing intensity of D_2 strain from N to S. L_1 shows progressive rotation from an N-plunging orientation towards an EW-orientation from the northern to the central part of the island. The oblique sinistral shear component is shown by shear bands at Stinker Point (inset); compare also with Fig. 5.

Fig. 7. D_3 kink bands showing late, post-metamorphic deformation.

therefore, were interpreted to have grown during D_2 [see fig. 17 in Trouw et al. (1998b)].

Albite porphyroblasts generally contain inclusions of straight S_1, locally associated with tight D_1 folds, in the core, and D_2 folds in the rim [see fig. 5 in Trouw et al. (1991); see also figs. 11.17 and 11.18 in Passchier and Trouw (1996)]. This means that the cores of these crystals grew intertectonically, after D_1 and before D_2 (Passchier and Trouw, 1996) and the rims during or after D_2. The transition between cores and rims is relatively abrupt; this is interpreted to indicate that albite growth continued during and after a relatively sudden initiation of D_2 folding. The relation between the growth of other metamorphic minerals

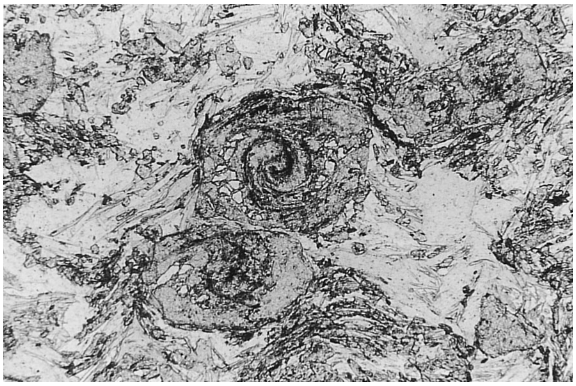


Fig. 8. Photomicrograph of garnet with spiral-shaped inclusion pattern, grown syntectonically to D_1, indicative of tectonic movement with top to the NE. Plane polarised light. Width of view 3 mm.

and the deformation phases is discussed by Trouw et al. (1998a,b) and is shown in Table 1.

6. Structural data from adjacent islands

Other islands of the Elephant Island group where the Scotia metamorphic complex crops out are Clarence Island, Gibbs Island and the group of Eadie, Aspland and O'Brien islands. A review of structural data from Smith Island and the western South Orkney Islands is also given for the sake of comparison.

6.1. Clarence Island

Hervé and coworkers (Hervé et al. 1983; Hervé and Pankhurst, 1984), Dalziel (1984), Marsh and Thompson (1985) and Trouw (1988) published structural data from this island. S_0–S_1 is close to horizontal or slightly west-dipping on the eastern part of the island and steep, with NE–SW strike along the west coast. The L_1 elongation lineation is about E–W and close to horizontal. D_2 folds have NE–SW-striking axial surfaces with shallow axes in the same direction. D_3 structures are restricted to local kink folds with irregular orientation.

6.2. Gibbs Island

De Wit et al. (1977) described the structures of this island in detail [see also comments by Dalziel (1984) and Trouw (1988)]. The Gibbs Island slide zone that separates a large dunite/serpentinite body from underlying schists has a WNW strike with moderate dip to the south. We interpret this slide zone as a D_1 structure, because of its parallelism with the main cleavage, S_1. A few elongation lineations with a shallow to moderate dip to the west were reported by De Wit et al. (1977) and we measured southwest dipping lineations related to the slide zone. D_2 deformation seems to have been more restricted at this island, possibly because of the different rheology of the dunite body. D_3 structures are similar to those described on Elephant Island.

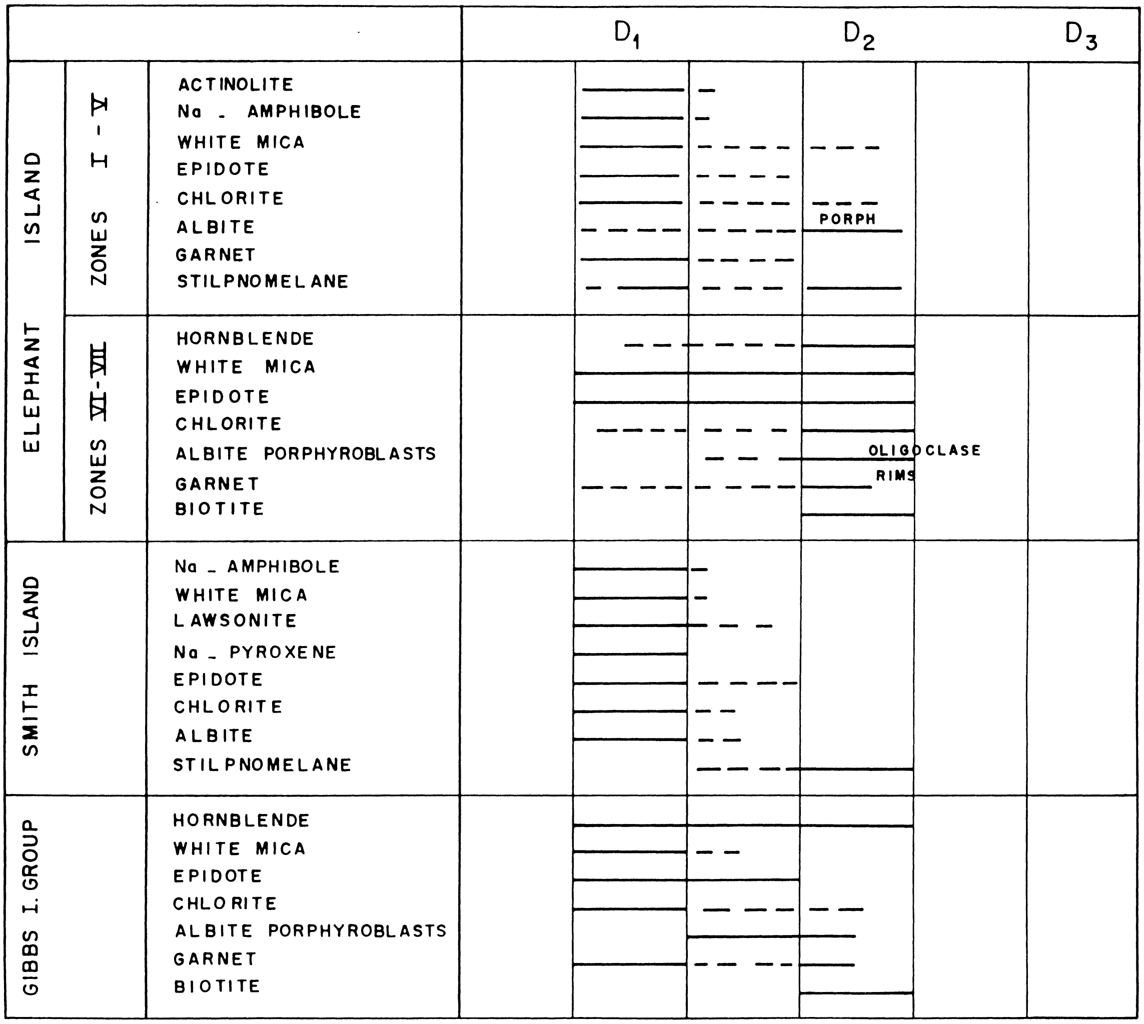
6.3. Aspland, Eadie and O'Brien islands

The main schistosity, interpreted as S_1, dips either south (Aspland) or east (Eadie and O'Brien) and is strongly folded. L_1 stretching lineations plunge moderately to the ESE. Dalziel (1984) attributed the folds to two conjugate sets, both related to his late phase, but we prefer to interpret them as D_2 folds, since thin section studies failed to reveal any older tectonic structure than the folded S_1. Axial planes of D_2 folds are steeply south dipping (Aspland) or steeply NE dipping (Eadie), with moderately to steeply SW- or SE-plunging axes.

6.4. Smith Island

Structural data from this island were presented by Dalziel (1984), Trouw (1988) and Grunow et al. (1992). As on southern Elephant Island, intense D_2 deformation produced transposition of earlier surfaces, but at Cape Smith moderately W- to NW-dipping enveloping surfaces of $S_1 \| S_0$ are recognisable. S_2 planes dip about 60° NW, whereas D_2 fold axes and mineral/extension lineations plunge moderately to the NE. D_3 structures are similar to those described for the other islands.

Table 1
Relationship between growth periods of metamorphic minerals and deformation phases; porph: porphyroblasts



6.5. *The western South Orkney Islands*

The structural evolution of the Scotia metamorphic complex outcropping on the South Orkney Islands was described by Dalziel (1984), Meneilly and Storey (1986) and Trouw et al. (1997). According to these last authors, five deformational phases were recognised at Powell Island. The main foliation, S_2, dips shallowly towards the south with SSE-plunging stretching lineations. D_3 produced folding of S_2 around axes with similar orientation. Later extensional shear bands with down dip movement in the same direction were attributed to D_4, and D_5 is a phase of brittle faults and kink bands. Meneilly and Storey (1986) reported a similar sequence of five deformation phases from Signy Island, interpreted as related to relative tectonic transport to the north or NNW as a consequence of subduction in the opposite direction.

7. Discussion

A first point to be discussed is the kinematic and tectonic significance of D_1 and D_2. Do these phases reflect different stages in a continuous deformation with similar kinematics or do they represent different kinematics? Available radiometric ages (Grunow et al., 1992) do not show an age gap between minerals grown during D_1 and D_2; therefore, they are considered as probably continuous in time. However, the kinematics of the phases are quite different, as discussed below.

7.1. *Kinematic and tectonic significance of D_1*

The fact that D_1 structures are the oldest tectonic features in the rocks that crop out at Elephant Island, and the parallelism of syntectonic glaucophane and crossite crystals with S_1 and L_1 structures was interpreted (Trouw, 1988; Trouw et al., 1998a) to indicate that D_1 acted contemporaneously with relatively high-pressure subduction-related metamorphism and that D_1 therefore reflects subduction movements. The presence of D_1-sheath folds (Trouw et al., 1991) and syn-D_1 garnets with spiral-shaped inclusions supports this interpretation. Therefore, D_1 structures provide information related to the subduction process during the Cretaceous. It is assumed here that S_1 formed originally with its strike approximately parallel to the direction of the trench and that L_1, interpreted as the direction of the main tectonic transport, was subparallel to the direction of subduction. It is obviously difficult to determine absolute directions of tectonic transport in the absence of palaeomagnetic data, but, where D_2 and D_3 are weak, it is possible to define a subduction direction relative to the present geographic reference frame. This is the case at Point Wild. We therefore interpret the E–W strike of S_1, combined with the steeply north-plunging attitudes of L_1 at Point Wild, as indicative of N–S subduction movements along an E–W-trending trench, in the present geographic reference frame. In southward subduction the lineations would be expected to plunge to the south as well. The fact that they plunge to the north can be explained as the result of large-scale D_2 folding around E–W axes or as an original inhomogeneity in the D_1 strain pattern. The garnets with spiral inclusion patterns, syntectonic with respect to D_1 (Fig. 5; Trouw, 1988; Trouw et al., 1991), indicate relative top-to-the-north shear movements, assuming that they rotated with respect to the surrounding cleavage, S_1. The sites where L_1 has an E–W attitude with low rake generally coincide with areas where D_2 overprinting is relatively strong. We therefore deduce that during D_2 the north- (or south-) plunging L_1 lineations were progressively rotated to an E–W position with lower rake. The fact that the girdle of L_1 attitudes lies slightly clockwise from the girdle of L_2 orientations (Figs. 3 and 4) is interpreted as being the result of the same progressive rotation by D_2. Sinistral D_2 non-coaxial flow along steep E–W shear zones, would reorient north-plunging L_1 lineations to an NE–SW orientation at low D_2 strain, and to an almost E–W orientation at high D_2 strain. The angle between the girdles of L_2 and L_1 (Fig. 3a and c) is in agreement with a sinistral rotation of L_1 and L_2 towards the fabric attractor (Passchier, 1997) in vertical, approximately E–W-striking, D_2 shear zones.

The outcrop pattern of the metamorphic isograds (Fig. 2a) and their polarity provide independent indicators of the orientation of the palaeotrench and subduction direction. Ernst (1975) demonstrated that the metamorphic polarity (that is the vector from lower to higher grade) usually indicates the direction of subduction. At Elephant Island this would mean subduction towards the SW. This method, however, is only capable of determining the subduction direction to a first approximation, since it does not take oblique subduction into consideration. One might question, in this context, why the isograds are oblique to the main structural trend. We consider that the restricted size of Elephant Island does not permit one to establish the regional importance of this obliquity; it might simply reflect a local irregularity related to the shape of the colliding plateau (see below).

Another independent confirmation of the general sense of subduction can be derived from the fact that the Cretaceous magmatic arc is located along the Antarctic Peninsula, south of Elephant Island (Leat et al., 1998).

An obvious problem in the determination of the 'original' subduction direction, is that of finding a suitable reference frame; both internal ductile deformation and rigid body rotation of the rock units underlying the island may have modified the orientation of lineations in a geographical reference frame. If we use the north coast of the island as our reference axis, we infer that L_1 in the north rotated little with respect to this axis, but L_1 in the centre may have rotated by up to 90° during D_2 (Figs. 3 and 4).

In order to assess the subduction direction on a regional scale, it is necessary to determine the orientation of L_1 with respect to geographical coordinates during D_1. D_1 was dated in the range 90 to 100 Ma (Trouw et al., 1990; Grunow et al., 1992), and available reconstructions of the orientation of the Pacific margin segment in which Elephant Island was situated (e.g. Barker et al., 1991; Grunow et al., 1992; McCarron and Larter, 1998) imply that the orientation of the north coast of the island may have been approximately NE–SW at that time. If L_1 did not rotate with respect to the north coast, subduction was accordingly approximately towards the southeast. This is in agreement with the general geographic setting with the proto-Pacific ocean in the west and with regional reconstructions for the period considered (e.g. Larson and Pitman, 1972; Zonenshayn et al., 1984; Scheuber and Andriessen, 1990). It is therefore assumed that at least the northern part of the island rotated in a rigid fashion after D_1 (and D_2), whereas the southern part was affected by both a rigid rotation and reorientation of D_1 structures during ductile D_2 deformation (Fig. 6).

7.2. *Kinematic and tectonic significance of D_2*

D_2 was interpreted by Trouw (1988) as a phase of horizontal shortening resulting in relative obduction of the rock units now outcropping at Elephant Island towards the SSE. The main argument for this 'obduction' is the asymmetry of D_2 folds with vergence towards the south (Fig. 2a), contrary to what would be expected in an accretionary wedge with subduction towards the south. With the new data presented here, it is now clear that D_2 is actually a phase of non-coaxial shear with an important oblique sinistral displacement component.

The orientation of S_2 planes, L_2 lineations and the C′-type shear bands fit a model of D_2 flow as constrained in one or more steep WSW–ENE-oriented shear zones (Figs. 5 and 6) with sinistral and 'north-downwards' movement components.

The intense folding of older fabric elements, with axes parallel to L_2, is one of the most striking features of D_2 deformation on Elephant Island. If D_2 flow was exclusively accomplished by simple shear, folding on this scale would only be expected where the older foliation was considerably oblique to the (steep) D_2 flow plane. Among other possible positions, this would be the case where S_0 and S_1 were gently dipping at the onset of D_2. A situation like that would lead to an oblique relation between L_2 and D_2 fold axes, especially in low-strain sections of the shear zone. However, L_2 is parallel to D_2 fold axes wherever measured. It is therefore probable that D_2 flow was not simple shear, but that it deviated from plane strain with an N–S-shortening component, i.e. a transpressional D_2 flow type. In such a flow type, abundant folds with axes parallel to the stretching lineation could develop and L_2 would be generated parallel to the D_2 fold axes, even in low strain domains.

The present D_2 structure of the whole island configures an irregular asymmetric dome-shape with lineations along the west coast dipping to the west and, at several places on the east coast (e.g. Hut Bluff and Walker Point), to the east (Fig. 6). This dome shape could be the result of an inhomogeneous secondary vertical flow component during transpression. The NNW–SSE-shortening component of D_2 is mainly suggested by the dominance of tight D_2 folds with steep axial planes and gently plunging axes, parallel to L_2. As stated above, the intensity of folding on this scale is not likely to be the result of simple shear alone. Since there are no indications for massive volume change during D_2, the shortening component must have been compensated by extension in other directions. Although vertical extension is usually assumed in transpression models (Harland, 1971; Sanderson and Marchini, 1984; Means, 1989; Fossen and Tikoff, 1993; Krantz, 1995), there are several indications that an important horizontal extension component was present during D_2 on Elephant Island.

(1) Strong D_2 C′ shear band cleavage is present with steep intersection lineations between C′ and S planes. Theoretical, experimental and field-oriented research (Passchier, 1991) has shown, that C′-type shear band cleavage is best developed in 'stretching' shear zones (Means, 1989), i.e. those with an extension component parallel to the direction of tectonic transport. (2) The orientation of the strong L_2 stretching lineation and the fact that the D_2 folds are strongly cylindrical parallel to L_2. If extension during D_2 were to have been mainly vertical, folds with more irregularly oriented axes should have been formed.

The mean value of the L_2 plunge, interpreted as the principal movement direction, is about 30° to the WSW, indicating that the strike-slip component of movement was about twice as important as the dip-slip component. If D_2 deformation is interpreted as related to the arrival of thickened oceanic crust in a subduction setting (Trouw et al., 1998a), then this strike-slip component could be the result of oblique subduction, possibly as a deep-seated equivalent to trench-linked strike-slip faults (Woodcock and Fischer, 1986; Sylvester, 1988) in shallower crustal levels. Although these faults generally occur within the magmatic arc (e.g. Hla, 1987; Scheuber and Andriessen, 1990), at a distance from the trench, contrary to the case considered here, their tectonic significance could be equivalent. It is worth noting, however, that at shallow levels, where brittle deformation predominates, the oblique transpressive movement may be strongly partitioned between subduction and strike-slip faults, whereas at a depth where ductile deformation prevails, the movement is expected to be resolved along a unique transpressive oblique shear zone, although some studies suggest the opposite (Hollister and Andronicos, 1997; Andronicos et al., 1999). A problem with this explanation is that published restorations for the regional tectonic setting of Elephant Island in the mid-Cretaceous (Barker et al., 1991; Grunow et al., 1992; McCarron and Larter, 1998) show the island in a position in relation to the trench and plate movement vectors, which would lead to dextral rather than sinistral oblique subduction.

An alternative explanation for the D_2 deformation with its sinistral strike-slip component at Elephant Island is to relate it to deformation within the forearc (Grunow et al., 1992), resulting from transcurrent motion between the Antarctic Peninsula and southern South America, initiated in the mid-Cretaceous. In this interpretation the dip-slip component of motion can only be explained by later tilting of the whole block, of which Elephant Island is the now outcropping area. However, this tilting of originally horizontal L_2 lineations would lead to deeper erosion of the eastern part of the island with consequently higher metamorphic grade. The metamorphic isograd pattern (Fig. 2a) shows quite the opposite, so this explanation seems problematic as well.

We conclude that D_2 was a phase of sinistral transpressive shear that, contrary to D_1, brought the considered rock units into a lower lithostatic pressure environment (Figs. 6 and 10). The meta-

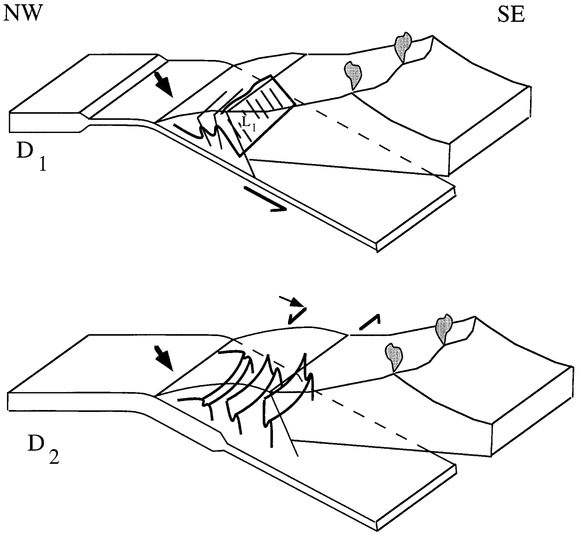


Fig. 9. Tentative scheme of tectonic evolution of Elephant Island during D_1 and D_2. During D_1, subduction towards the S or SE led to steep S- to SE-plunging L_1 lineations and growth of relatively high-pressure low-temperature metamorphic minerals. Folding and sinistral transpressional movement during D_2 possibly resulted from the arrival of thickened oceanic crust during D_2 and associated minor collision. The steeply plunging L_1 lineations were folded to moderately SW-plunging orientations in the northern part of the island, and rotated to NE–SW positions in the remaining part of the island. Metamorphism changed to higher- temperature lower-pressure conditions.

morphic evolution during D_2 corroborates this interpretation, because pressures were falling and temperatures increasing (Trouw et al., 1998a), generating a classical clockwise *P–T–t* path (England and Thompson, 1984; Thompson and England, 1984).

The decrease of pressure during D_2 may have been caused by the vertical dislocation component of D_2 tectonic transport, indicated by the plunging nature of the L_2 stretching lineations, consistent with relative uplift of the southern part with respect to the northern part of the island. Alternatively, uplift may have been produced by a vertical extension component to movement inside the transpressive D_2 shear zone, causing secondary vertical expulsion of material in addition to the main movement along the stretching lineations. On a wider scale, emergence of the whole metamorphic pile may have been due to isostatic uplift after local stagnation or slowing down of subduction (Fig. 9).

7.3. Tectonic significance of D_3

As stated in Section 4.3 on D_3 structures, most kink bands were probably generated as a consequence of E–W shortening and N–S extension. Since D_1 and D_2 correspond to a large extent to N–S compression, D_3 can be interpreted as a phase of relaxation, related to uplift and exhumation. However, no low-angle normal faults or jumps in metamorphic grade were observed to indicate exhumation by extension. Isostatic uplift followed by erosion seems the most probable mechanism for exhumation. It is clear from the Cenozoic evolution of the Scotia Sea (Barker et al., 1991; Cunningham et al., 1995) that Elephant Island must have rotated over as much as 45° in a clockwise sense during the last 40 Ma. The E–W shortening and N–S extension are in good agreement with the strain orientation along the sinistral Shackleton Fracture Zone, active during the period 29–4 Ma (Klepeis and Lawver, 1996; Kim et al., 1997; Figs. 1 and 10). It is therefore concluded that D_3 structures were probably generated during the same time interval, and may be related to movement along the Shackleton Fracture Zone, associated with the opening of the Scotia Sea (Fig. 10).

7.4. Comparison with other islands

Although the orientation of the structures at Clarence Island is somewhat different from Elephant Island, the flat-lying foliations can be compared to the situation at central west Elephant

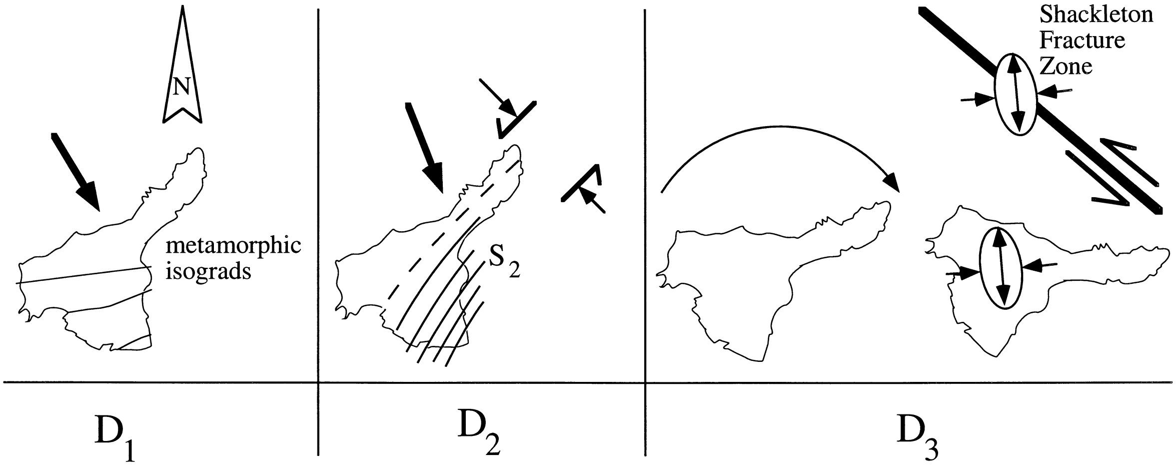


Fig. 10. Evolutionary scheme showing the original orientation of D_1 and D_2 structures on Elephant Island with respect to north, and their rotation, probably during D_3, related to the development of the Scotia arc and the Shackleton Fracture Zone.

Island and the shallow NE-dipping lineations to similar ones at Walker Point (Fig. 2b). The SSW-dipping D_1 slide zone at Gibbs Island with its W- to SW-plunging lineations might reflect southwestward subduction or southward subduction with superposed clockwise rigid body rotation. The south- to east-dipping foliations with ESE-plunging lineations on Aspland, Eadie and O'Brien islands are quite different from Elephant Island. At present it is not clear whether this might be due to block rotation, to large-scale folding or to another mechanism.

The orientation of structures at Smith Island, probably generated at about 50 Ma, is also different from Elephant Island in the sense that the lineations plunge to the ENE. It seems quite possible that these structures were generated by eastward oblique subduction with a dextral strike slip component, but no data to confirm this are presently available.

The orientation of structures on the South Orkney Islands led Meneilly and Storey (1986) and Trouw et al. (1997) to infer southward subduction with respect to the present orientation of these islands. Although metamorphic ages from these islands are in the range 180–200 Ma, almost twice as old as the ones from Elephant Island, the tectonic setting seems to be comparable to the situation during D_1 at Elephant Island.

8. Conclusions

The predominantly E–W strike of S_1 with N–S-oriented stretching lineations led to the interpretation that D_1 reflects southward subduction, with respect to the present orientation of Elephant Island.

D_2 modified the orientation of L_1 stretching lineations to an E–W position; this phase is interpreted as a phase of transpressional movement with a sinistral strike slip component, either produced by oblique subduction or by transcurrent motion between South America and the Antarctic Peninsula, in the fore-arc. Increasing temperature and decreasing pressure during this phase probably result from the slowing down of subduction movements due to the arrival of thickened oceanic crust, which led to the northward migration of the active subduction zone.

Post-metamorphic D_3 kink bands and local faults were generated during clockwise rigid body rotation over about 45° associated with uplift in the Cenozoic.

Comparison with other parts of the Scotia metamorphic complex shows comparable structural evolutions but often considerable differences in orientation.

Acknowledgements

The data presented in this paper were collected during two Brazilian expeditions in the austral summers of 1991–1992 and 1995–1996. Mônica Heilbron and Julio César de Almeida are gratefully acknowledged for enthusiastic assistance in the field and in the laboratory. This research was financed by CNPq PROANTAR, grant 404513. CWP acknowledges financial support from the 'Stichting GOA'. Constructive reviews by Keith Klepeis and an anonymous reviewer are gratefully acknowledged.

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