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

Rudolph A.J. Trouw a,*, Cees W. Passchier b, Claudio M. Valeriano c, Luiz Sérgio A. Simões d, Fabio V.P. Paciullo a, André Ribeiro a

a Departamento de Geologia, I.GEO, UFRJ, 21910-900 Rio de Janeiro, Brazil
b Institut für Geowissenschaften, Gutenberg Universität, 55099, Mainz, Germany
c Departamento de Geologia/Geofísica, UERJ, 20559-900, Rio de Janeiro, Brazil
d Departamento de Petrologia/Metalogenia, UNESP, 13500-230, Rio Claro, SP Brazil

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, D1, produced a strong SL fabric with steep stretching and mineral lineations, partly defined by relatively high pressure minerals, such as crossite and glaucophane. D1 is interpreted to record southward subduction along an E–W trench with respect to the present position of the island. A second phase, D2, 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 L2 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 D2, interpreted to represent the collision of thickened oceanic crust with the active continental margin. The last deformation phase, D3, 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 D3, in Cenozoic times, together with the trench, from an NE–SW position, during the progressive opening of the Scotia Sea. The similarity between the strain orientation of D3 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

* Corresponding author. Fax: +55-21-5983280.
E-mail addresses: trouw@igeo.ufrj.br (R.A.J. Trouw), cpaschi@mail.uni-mainz.de (C.W. Passchier), cmval@vmesa.uerj.br (C.M. Valeriano), dp@geo001.uesp.ansp.br (L.S.A. Simões)

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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-day 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 (hemipelagic 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.
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 northeastern 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 \((\sim 7 \text{ kbar, } 350^\circ \text{C}; \text{Trouw et al., 1998a})\) related to \( D_1 \) is attributed to a subduction setting because of the corresponding low thermal gradient of about \( 17^\circ/\text{km} \). The higher temperature and somewhat lower pressure conditions, during \( D_2 \) \((\sim 5 \text{ kbar, } 500^\circ \text{C})\), 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
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 metacon-
glomerates 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 elon-
gate 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 develop-
ment 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

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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.
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.

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
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 $S_1 \parallel S_0$ are recognisable. $S_1$ 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

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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, S₂, dips shallowly towards the south with SSE-plunging stretching lineations. D₃ produced folding of S₂ around axes with similar orientation. Later extensional shear bands with down dip movement in the same direction were attributed to D₄, and D₅ 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 D₁ and D₂. 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 D1 and D2; 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 D1

The fact that D1 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 S1 and L1 structures was interpreted (Trouw, 1988; Trouw et al., 1998a) to indicate that D1 acted contemporaneously with relatively high-pressure subduction-related metamorphism and that D1 therefore reflects subduction movements. The presence of D1-sheath folds (Trouw et al., 1991) and syn-D1 garnets with spiral-shaped inclusions supports this interpretation. Therefore, D1 structures provide information related to the subduction process during the Cretaceous. It is assumed here that S1 formed originally with its strike approximately parallel to the direction of the trench and that L1, 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 D2 and D3 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 S1, combined with the steeply north-plunging attitudes of L1 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 D2 folding around E–W axes or as an original inhomogeneity in the D1 strain pattern. The garnets with spiral inclusion patterns, syntectonic with respect to D1 (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, S1. The sites where L1 has an E–W attitude with low rake generally coincide with areas where D2 overprinting is relatively strong. We therefore deduce that during D2 the north- (or south-) plunging L1 lineations were progressively rotated to an E–W position with lower rake. The fact that the girdle of L1 attitudes lies slightly clockwise from the girdle of L3 orientations (Figs. 3 and 4) is interpreted as being the result of the same progressive rotation by D2. Sinistral D2 non-coaxial flow along steep E–W shear zones, would reorient north-plunging L1 lineations to an NE–SW orientation at low D2 strain, and to an almost E–W orientation at high D2 strain. The angle between the girdles of L2 and L1 (Fig. 3a and c) is in agreement with a sinistral rotation of L1 and L2 towards the fabric attractor (Passchier, 1997) in vertical, approximately E–W-striking, D2 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 \( \mathbf{L}_1 \) in the north rotated little with respect to this axis, but \( \mathbf{L}_1 \) 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 \( \mathbf{L}_1 \) 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 \( \mathbf{L}_1 \) 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; Zonenshain 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 D₂ 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 L₂ stretching lineation and the fact that the D₂ folds are strongly cylindrical parallel to L₂. If extension during D₂ were to have been mainly vertical, folds with more irregularly oriented axes should have been formed.

The mean value of the L₂ 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 D₂ 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 D₂ 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 L₂ 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 D₂ was a phase of sinistral transpressive shear that, contrary to D₁, 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 D₁ and D₂. During D₁, subduction towards the S or SE led to steep S- to SE-plunging L₁ lineations and growth of relatively high-pressure low-temperature metamorphic minerals. Folding and sinistral transpressional movement during D₂ possibly resulted from the arrival of thickened oceanic crust during D₂ and associated minor collision. The steeply plunging L₂ 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.](image-url)
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

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**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 transtensive 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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