Mesozoic tectonic evolution of the South Orkney Microcontinent, Scotia arc, Antarctica

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Abstract – The South Orkney Islands are the exposed part of a continental fragment on the southern limb of the Scotia arc. The islands are to a large extent composed of metapelites and metagreywackes of probable Triassic sedimentary age. Deformation related to an accretionary wedge setting, with associated metamorphism from anchizone to the greenschist facies, are of Jurassic age (176–200 Ma). On Powell Island, in the centre of the archipelago, five phases of deformation are recognized. The first three, associated with the main metamorphism, are tentatively correlated with early Jurassic subduction along the Pacific margin of Gondwana. D_4 is a phase of middle to late Jurassic crustal extension associated with uplift. This extension phase may be related to opening of the Rocas Verdes basin in southern Chile, associated with the breakup of Gondwanaland. Upper Jurassic conglomerates cover the metamorphic rocks unconformably. D_5 is a phase of brittle extensional faulting probably associated with Cenozoic opening of the Powell basin west of the archipelago, and with development of the Scotia arc.

1. Introduction

The South Orkney Islands are located at 60°35'S, 45°00'W on the south side of the Scotia arc, between South America and the Antarctic Peninsula (Fig. 1). The islands are the only exposed part of a microcontinent known as the South Orkney Microcontinent (SOM) (Dalziel, 1984; Barker, Dalziel & Storey, 1991). The SOM is one of several continental fragments that form both the North and South Scotia ridges (Fig. 1) (Dalziel, 1984; King & Barker, 1988; Barker, Dalziel & Storey, 1991; King *et al.* 1994). These fragments are part of a once continuous active plate margin of Gondwanaland and were brought into their present position by differential relative motion of southern South America and the Antarctic Peninsula (Cunningham *et al*. 1995) and by back-arc spreading in the Scotia arc within the past 40 Ma (Barker, Dalziel & Storey, 1991). This spreading formed the Scotia and Sandwich plates and seems to have occurred, during the last 8 Ma before present, in response to rapid eastward roll-back of the hinge of a subduction zone that preceded the present South Sandwich arc (Fig. 1); active spreading is still taking place behind this arc in response to westward subduction of the South American plate (Barker, Dalziel & Storey, 1991). The SOM is separated from the Scotia plate in the north by an active sinistral transform fault; in the west and east it is bounded by small extensional oceanic basins, the Powell and Jane basins, and in the south by the Weddell basin (Barker, Barber & King, 1984; King & Barker, 1988; Lawver, della Vedova & von Herzen, 1991; Barker, Dalziel & Storey, 1991). The southwest margin of the SOM may be a strike-slip fault parallel to the sinistral transform boundary to the north (King & Barker, 1988). A narrow ridge probably of extended continental crust, the South Scotia Ridge, connects the SOM with the continental crust of the Antarctic Peninsula (Galindo-Zaldivar *et al.* 1994).

Although the SOM is relatively small and only one of many continental fragments in the Scotia Arc region, its tectonic history is of more than local interest. The South Orkney Islands are mainly composed of metasedimentary and meta-igneous rocks that accumulated on the Pacific margin of Gondwanaland prior to Gondwana breakup, and little is known about the tectonic regime that existed during this period in West Antarctica. Gondwana breakup, initiated in this region by the separation of an Andean/Antarctic Peninsula magmatic arc from the remainder of South America, during late Jurassic and early Cretaceous times, led to the opening of the Weddell sea (Barker, Dalziel & Storey, 1991; King, Livermore & Storey, 1996; DiVenere, Kent & Dalziel, 1996). Finally, the Scotia arc formed a continuous connection between Antarctica and South America after Gondwana breakup until *c.* 30 Ma, and prevented establishment of the circumpolar current with its consequences for the climate in the southern hemisphere until that time (Barker, Dalziel & Storey, 1991). The connection disintegrated during initial stages of the development of active spreading centres in the western Scotia Sea (Barker & Burrell, 1977; Cunningham *et al.* 1995), but details of this breakup are not well known. Since the SOM was situated in the region where disintegration started, its tectonic history may help to clarify the evolution of the Scotia arc as a whole.

2. Regional geology

The South Orkney Islands are composed of five major and a large number of smaller islands. The principal islands are Coronation, Powell, Fredriksen, Laurie and Signy (Fig. 1b). The islands have a relatively steep relief with a maximum elevation of about 1000 m, and nearly 90 % of the surface is covered by glaciers.

The islands are composed of three major rock units (Fig. 1b) (Thomson, 1973, 1974; Dalziel, 1984): (i) the metamorphic complex, underlying Coronation, Signy and northern Powell islands; (ii) the Greywacke-Shale Formation (Thomson, 1973) cropping out on Laurie and surrounding islands, and at southern Powell Island; and (iii) the Spence Harbour and Powell Island conglomerates (Thomson, 1973; Elliot & Wells, 1982; Wells, 1984).

The metamorphic rocks are part of the Scotia metamorphic complex (Tanner, Pankhurst & Hyden, 1982; Dalziel, 1982, 1984) that has been interpreted as a Mesozoic–Cenozoic subduction complex generated by the subduction of Pacific Ocean floor (Smellie & Clarkson, 1975; Dalziel, 1984; Meneilly & Storey, 1986). The complex is defined by metamorphic criteria and may therefore include rocks of different age and protolith. At the South Orkney Islands, radiometric age dating (K–Ar, Rb–Sr) indicates that the metamorphism occurred between 205 and 176 Ma (early to middle Jurassic: Miller, 1960; Grikurov, Krylov & Silin, 1967; Tanner, Pankhurst & Hyden, 1982; Grunow *et al.* 1992).

The Greywacke-Shale Formation is mainly composed of stratified successions, interpreted as submarine fan deposits resulting from turbidity current processes. This formation is considered to be an equivalent of the Trinity Peninsula Group from the Antarctic Peninsula (e.g. Dalziel, 1984), with similar lithofacies and environmental interpretation (Smellie, 1991; Paciullo, Andreis & Ribeiro, 1994; Ribeiro *et al.* 1994). The Trinity Peninsula Group and its correlatives have been ascribed a presumed Permo-Triassic age (Smellie, 1991), but fossil content (Thomson, 1975; Dalziel *et al.* 1981) and most Rb–Sr age determinations (Willan, Pankhurst & Hervé, 1994; Trouw, Pankhurst & Ribeiro, 1997) point to a Triassic rather than a Permian sedimentation age.

The Spence Harbour and Powell Island conglomerates cover both of the above mentioned units unconformably. They are almost entirely composed of coarse debris derived from these units and were interpreted as alluvial fan deposits (Elliot & Wells, 1982; Wells, 1984). Fossils indicate an early Cretaceous age for the Spence Harbour Conglomerate and a late Jurassic–early Cretaceous age for the Powell Island Conglomerate (Thomson, 1981).

Thomson (1973, 1974) published general descriptions of the geology of Powell and Coronation islands. Storey & Meneilly (1985) discussed the paragenesis of metamorphic rocks from Signy Island (Fig. 1b), including whole rock and mineral chemistry data, confirming that they were formed in a subduction complex setting. The tectonic transport recorded in the rocks of this island was deduced to be top towards the north–northwest along flat-lying foliation surfaces (Meneilly & Storey, 1986).

Dalziel (1984) described the structure of the whole archipelago in detail; he drew attention to the fact that the only known contact between the Scotia metamorphic complex and the pre-late Jurassic sedimentary sequences (Greywacke-Shale Formation, Trinity Peninsula Group) occurs on Powell Island. For this reason Powell Island was chosen for detailed study during the 8th and 10th Brazilian Expeditions to Antarctica, in the austral summers of 1989/90 and 1991/92. This paper presents new data on Powell and neighbouring islands, and discusses the implications for reconstructions of the Mesozoic tectonic evolution of the Scotia arc region. The data are derived from a total of 23 field stations shown in Figures 1b and 2. These stations were visited either by inflatable boat or by helicopter; at each station a detailed analysis of the structure was made and samples were collected.

3. Lithotypes and protoliths

3.a. Powell Island

The present account of the geology of Powell Island is mainly based on the analysis of outcrops along the coast. The topography of the 1:100 000 sheet (British Antarctic Survey, 1988) is inaccurate and was corrected where possible (Fig. 2). On previously published maps of Powell Island (e.g. Dalziel, 1984) all three rock units mentioned in the introduction are present (Fig. 1b), and this is, in a way, correct. However, our study confirms that a gradual metamorphic and structural transition occurs between the low grade Greywacke-Shale Formation in the south and higher grade ‘metamorphic complex’ rocks in the north. This possibility was already postulated by Thomson (1973) and by Dalziel (1984), who discussed the transition in detail. Dalziel (1984) reached the conclusion that it is a transition “(1) in a structural sense, the same foliation gradually becomes more penetrative northward into the metamorphic complex; (2) in a metamorphic sense, the grade and degree of recrystallization increase northward, and (3) also in a lithologic sense…”. In our understanding the lithological transition is only of local importance, as further explained in the next paragraph. This means that in our view the transition is a zone through which the Greywacke-Shale Formation rocks become gradually more deformed and metamorphosed towards the the north. It therefore does not justify a lithological contact on the map (Fig. 2).

At southern Powell Island (PO-1; Fig. 2) metasand- and siltstones appear more or less regularly interlayered

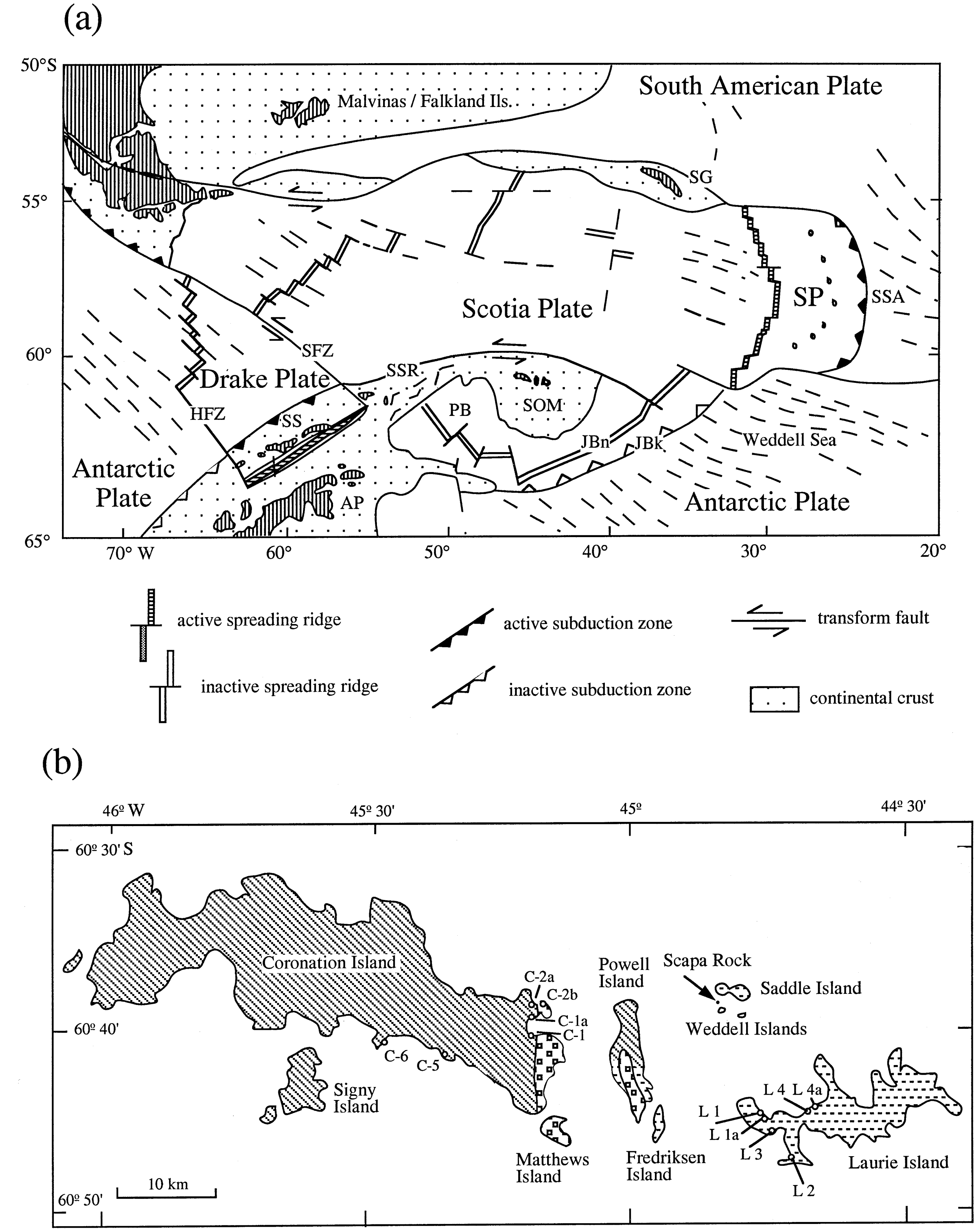

Figure 1. (a) Tectonic setting of the South Orkney Islands, modified after Galindo-Zaldivar *et al.* 1994. AP – Antarctic Peninsula; JBk – Jane Bank; JBn – Jane Basin; HFZ – Hero fracture zone; PB – Powell Basin; SFZ – Shackleton fracture zone; SG – South Georgia; SOM – South Orkney Microcontinent; SP – Sandwich plate; SS – South Shetland Islands; SSA – South Sandwich arc; SSR – South Shetland ridge. (b) Geological map of the South Orkney Islands (after Thomson, 1973; Dalziel, 1984). Diagonal traces: metamorphic complex; horizontal dashes: Greywacke-Shale Formation; blocks: Spence Harbour and Powell Island conglomerates. Numbered points are visited stations.

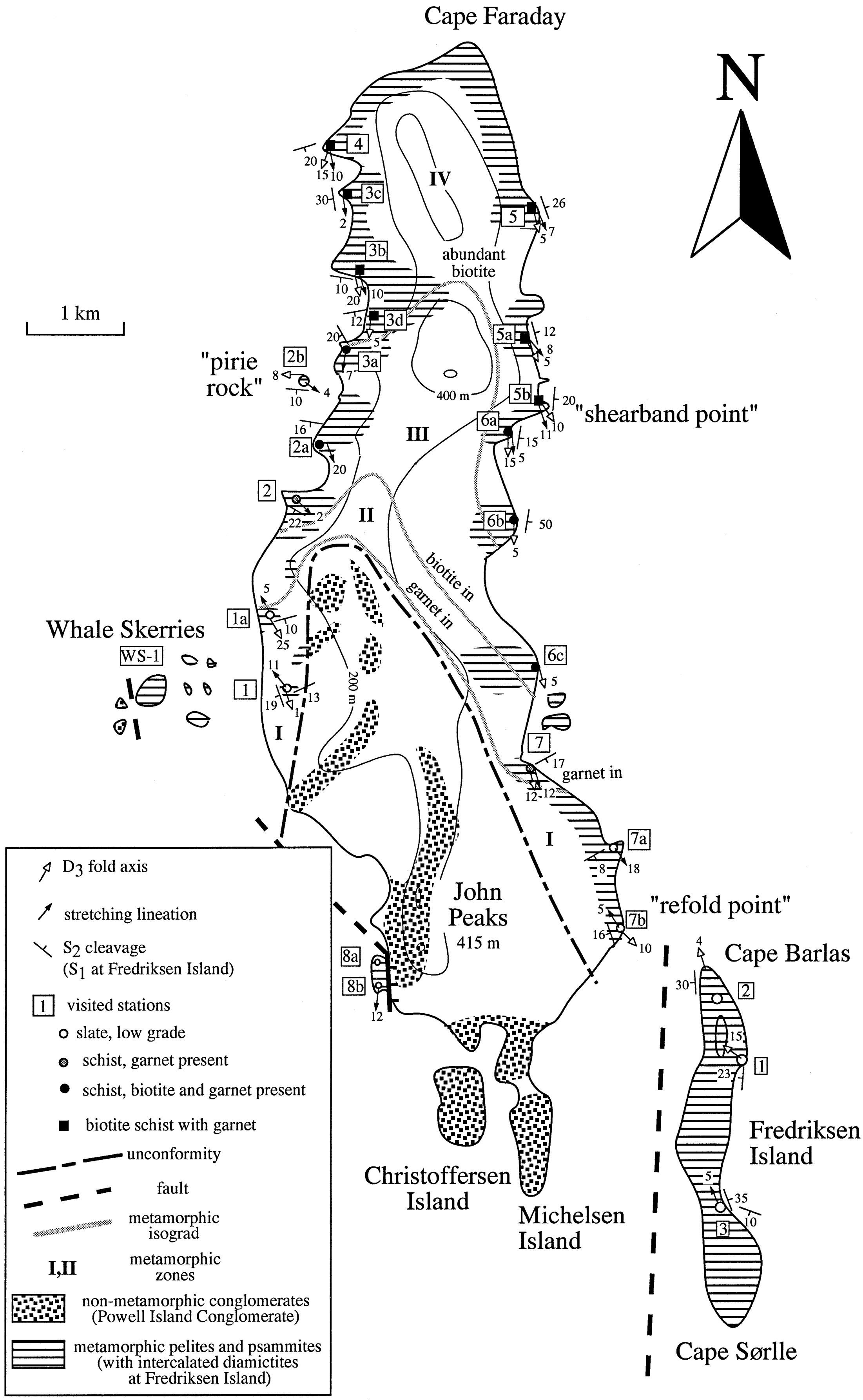

Figure 2. Geological map of Powell and surrounding islands showing visited stations and orientation of structures. Stations 1, 2 etc. at Powell Island are referred to in the text as PO-1, PO-2 etc. and at Fredriksen Island as FR-1, 2 etc.

in beds of 1 to 50 cm thickness. Intense deformation has destroyed most original sedimentary structures, but correlation with nearby Fredriksen Island (where these structures are better preserved) suggests that these rocks belong to a marine turbidite sequence as discussed in Section 3.c. Further to the north, bedding is progressively transposed by metamorphic and deformational processes and the metasand- and siltstones change gradually into grey phyllites and schists. These, studied in thin section, reveal a surprisingly homogeneous composition that can be summarized as follows: quartz 25–50 %, albite 25–50 %, white mica 10–25 %, chlorite 1–15 %, epidote 0–10 % and calcite 0–10 %. Common accessories are tourmaline, sphene and apatite. In addition, garnet and biotite appear (up to 5 % each) in the northern part of Powell Island. At some places (e.g. station PO-5) varieties richer in calcite (up to 25 %) are also present. At stations PO-3A, PO-5A and PO-6C (Figs 2, 3) a greenschist occurs as intercalations of up to a few metres in thickness. This greenschist locally contains large (up to 1.5 cm) garnet porphyroblasts and is further composed of the same minerals as the metasand- and siltstones, but in different proportions: albite 20–35 %, epidote 20–35 %, chlorite 15–25 %, garnet 1–10 %, calcite 2–6 %, sphene 2–10 % and quartz 1–5 %. Traces of green amphibole and biotite are locally present. The greenschist may be derived from (i) intrusive or extrusive mafic igneous rocks, (ii) marls or (iii) sandstones composed predominantly of mafic lithic fragments. Since transitional compositions between greenschist and metasandstones have been found at some outcrops, the last two options are most likely.

3.b. Whale Skerries

This group of islands west of Powell Island (Fig. 2) is mainly composed of Greywacke-Shale Formation rocks (Thomson, 1973). We only visited the main island, which was found to be essentially constituted of highly fractured and faulted massive sandstones and conglomerates with clastic fragments of up to about one centimetre in diameter. Sedimentary structures are generally destroyed by intense deformation.

3.c. Fredriksen Island

The lowest grade part of the sequence of Powell and surrounding islands crops out at Fredriksen Island, probably separated from southern Powell Island by an approximately north–south trending fault (Fig. 2). The existence of this fault is deduced from the considerable contrast in structural style and in intensity of metamorphic recrystallization between stations PO-7B and FR-2 (Fig. 2). We studied three stations on the island (Figs 2, 3). At station FR-2 a well-bedded (meta)sedimentary succession is preserved, mainly composed of slates (70 %) and feldspathic metasandstones (30 %) with at least three intercalations of diamictites (up to 6 m thick, Fig. 4). The fragments in these diamictites vary from small pebbles (less than one centimetre in diameter) to blocks of over a metre in diameter; most pebbles are composed of sandstone but at least one is a (sub)volcanic rock, although weakly metamorphosed.

Storey & Meneilly (1983) described a mélange at Fredriksen Island at a single visited station, "consisting of a chaotic arrangement of irregular sized blocks, up to 8 m across, of basic pillow lava, chert, felsite and epiclastic sandstone in a pervasively sheared cataclastic matrix". We visited this site as well (our station FR-1), but found that these blocks, studied in 22 thin sections, are mainly sandstone with some conglomeratic and few slightly metamorphosed lava pebbles. We interpret this outcrop as a deformed diamictite because we observed a planar contact in the northern part of this station with regular sandstone and pelite layers, similar to contacts observed along diamictites at station FR-2 (Fig. 4). At station FR-3 a similar, deformed diamictite crops out with pebbles and blocks of sandstone and slightly metamorphosed lava varying in size between a few centimetres and 10 m.

Detailed study of the sedimentary structures in the well-stratified part (Fig. 4) shows that the (meta)sedimentary sequence of Fredriksen Island represents part of a submarine fan composed of interlayered pelite and sandstone that results from turbidity current processes. Intercalations of diamictite facies are the result of debris flows. The interpretation of the diamictites as sedimentary debris flow deposits intercalated within the turbidite succession is in accordance with descriptions of similar

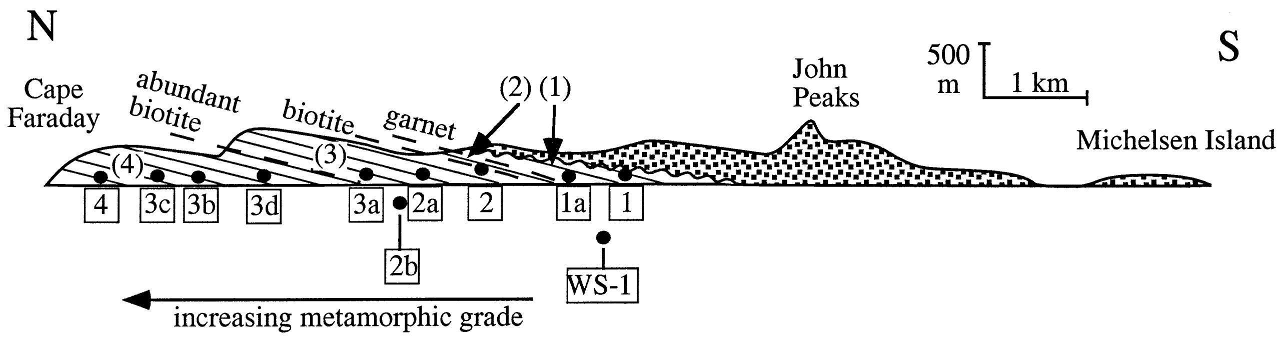

Figure 3. Geological section along the west coast of Powell Island; (1), (2), (3) and (4) refer to metamorphic zones (Fig. 2) explained in the text.

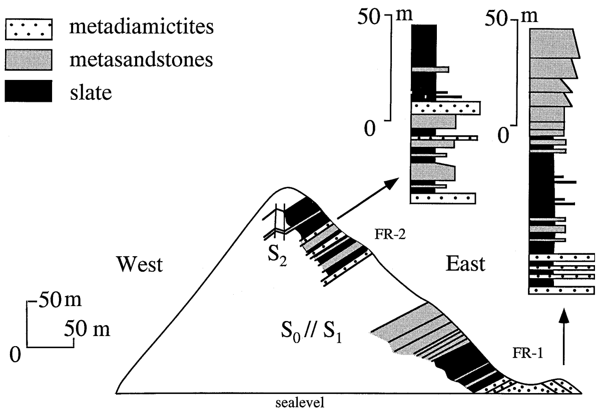

Figure 4. Geological section through Fredriksen Island with stratigraphic profiles elaborated at stations FR-1 and FR-2. Note the presence of meta-diamictite layers interstratified between metasandstone and slate.

occurrences in lithological units considered to be equivalent, such as the Trinity Peninsula Group at Cape Legoupil (Halpern, 1965; Ribeiro *et al.* 1994) and the Miers Bluff Formation at Livingston Island (Doktor, Swierczewska & Tokarski, 1994).

Table 1. Whole rock chemical composition of five volcanic pebbles from the Greywacke-Shale Formation at Fredriksen Island

	FR-1-8	FR-1-10	FR-2-12	FR-3-1	FR-3-2
SiO_2	48.60	48.90	52.00	49.40	47.00
TiO_2	1.90	0.66	2.40	2.10	2.30
Al_2O_3	13.90	15.80	15.40	15.20	16.60
Fe_2O_3	1.60	8.00	5.40	2.00	2.70
FeO	4.60	3.50	3.80	6.30	6.40
MnO	0.13	0.16	0.21	0.13	0.13
MgO	2.20	3.10	3.50	5.30	5.90
CaO	11.20	6.70	5.50	6.40	5.70
Na_2O	6.00	5.40	5.50	5.30	5.40
K_2O	0.17	0.13	0.10	0.09	0.35
P_2O_5	0.41	0.20	0.32	0.38	0.39
Total	90.71	92.55	94.13	92.60	92.87
Trace elements (ppm)					
Cr	68	890	68	233	226
Ni	-	-	-	110	110
Rb	11	10	8	12	20
Sr	270	240	350	340	528
Y	64	30	33	17	24
Zr	190	41	230	204	222
Nb	23.0	-	30.0	22.0	28.0
Ba	985	116	90	96	130
Co	36	45	48	-	-
V	169	281	225	-	-
Rare Earth Elements					
La	13.67	4.52	9.41	23.50	24.30
Ce	31.91	13.69	26.45	54.68	56.85
Nd	24.94	8.53	20.68	29.11	31.12
Sm	7.45	2.23	4.99	5.76	5.82
Eu	2.39	0.53	1.89	1.67	1.71
Gd	7.24	2.17	4.23	4.67	4.86
Dy	6.43	1.99	3.53	3.61	4.20
Ho	1.35	0.45	0.73	0.65	0.76
Er	3.38	1.52	2.17	1.39	1.67
Yb	2.11	1.12	1.69	0.75	0.99
Lu	0.25	0.15	0.22	0.10	0.10

For location of stations FR-1 to FR-3, see Figure 2.

The chemical compositions of the volcanic rock pebbles (Table 1) plot in the field of basaltic trachyandesites in the classification diagram of le Maitre (1989) (Fig. 5a) and show considerable differentiation in a chondrite normalized spidergram (Fig. 5b). The tendency of these rocks is alkaline rather than calc-alkaline and they do not define a clear tectonic environment. The analyses are somewhat different from analyses of mafic rocks from Signy Island (Fig. 1b) (Storey & Meneilly, 1985) that fall mainly in the basalt field in the classification diagram of le Maitre (1989). Storey & Meneilly (1985) interpreted the rocks from Signy Island as representing enriched tholeiitic and alkali basalts of an oceanic intraplate basalt series. In our understanding the composition of the pebbles is not conclusive and they may equally well be derived from a continental margin.

4. Provenance of metasandstones

Metasandstones from southern Powell Island (stations PO-8, PO-8A, PO-1), the Whale Skerries and Fredriksen Island (Fig. 2) can be classified as medium clean plagioclase arenites and less abundant wackes. Two populations were identified: (i) relatively quartz-rich arenites and (ii) arenites rich in plagioclase and lithic volcanic fragments. Detrital fragments have the following composition: quartz (monocrystalline > polycrystalline), feldspar (albite–oligoclase > orthoclase + microcline) and minor lithic components. In order of abundance the lithic components consist of: (i) clasts of volcanic/pyroclastic origin (including several types of andesites, vitric tuffs and andesitic ignimbrites); (ii) clasts of plutonic origin (tonalite, granodiorite > granite, monzonite); (iii) clasts of metamorphic rocks (phyllites, mica-schists, and less granite gneiss and mylonite); and (iv) clasts of sedimentary rocks (fine feldspathic arenites, siltstones and claystones). Most of these sedimentary rock fragments are restricted to coarse arenites and fine diamictites. Other clastic fragments in these arenites are: recrystallized volcanic glass (pecilite), lithic breccias (with abundant angular fragments of andesite), brown and partially chloritized biotite, chlorite, muscovite and accessory epidote, allanite, zircon and apatite. The less-deformed sandstones retain most of their sedimentary characteristics, such as clast outlines, that show mainly subangular to minor angular shapes. Circularity shows medium to high values (0.62–0.82). The contacts between clasts are mainly of the tangential and planar types, with less frequent concave–convex and sutured types; the latter become more frequent in more strongly deformed sandstones. In a few samples faint syntaxial growth of quartz (on quartz grains) and of albite on oligoclase grains was detected.

In the wackes ortho-matrix appears as recrystallized aggregates of white mica without preferred orientation. In the arenites, these aggregates appear as thin envelopes that partly or completely surround clasts. Reaction between these envelopes and clasts produces disappearance

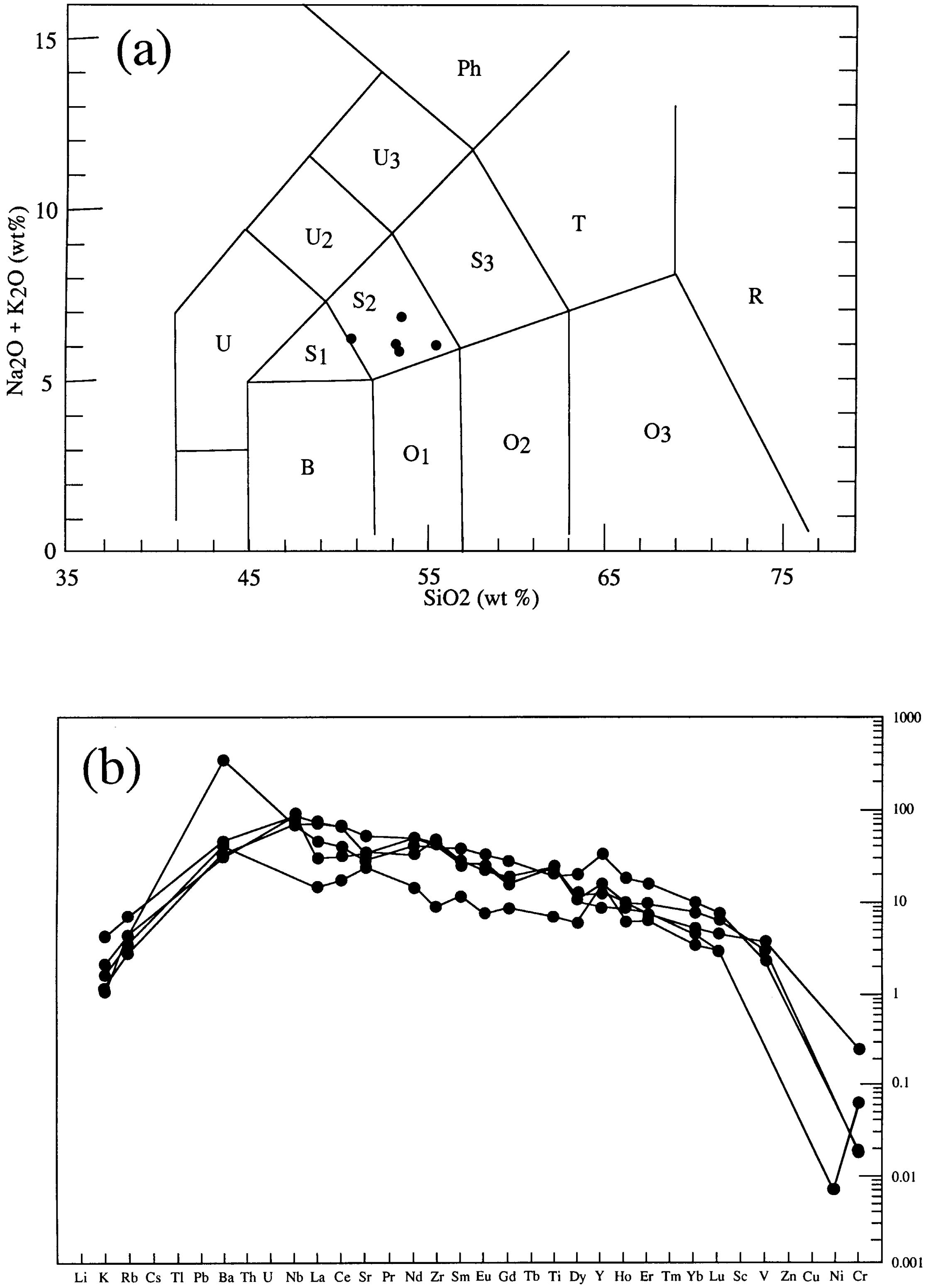

Figure 5. (a) Five analyses of volcanic lava pebbles from Fredriksen Island plotted in the compositional diagram of le Maitre (1989). B – basalt; O_1 – basaltic andesite; O_2 – andesite; O_3 – dacite; Ph – phonolite; R – rhyolite; S_1 – trachybasalt; S_2 – basaltic trachyandesite; S_3 – trachyandesite; T – trachyte and trachydacite; U – tefrite and basanite; U_2 – phonotephrite; U_3 – tephriphonolite. All analyses fall in field S_2 of basaltic trachyandesites. (b) Spidergram normalized to chondrite, showing considerable differentiation for the five analysed volcanic pebbles from Fredriksen Island.

of clastic borders. Except for the claystone intraclasts (pseudo-matrix), all other components are extrabasinal in origin.

The composition of the arenites is given in Table 2. In the Qt–F–L diagram (where L includes volcanic, sedimentary and metasedimentary fragments), given by Dickinson (1985) for tectonic classification of provenance (Fig. 6a), the samples plot mainly in the 'recycled orogen' field, and few in the 'dissected arc' field. In the Qm–F–Lt diagram (Dickinson, 1985) the sixteen samples are distributed in various fields (Fig. 6b): 'dissected arc' (5), 'quartzose recycled orogen' (5), 'mixed' (5) and 'transitional continental' (1). Valloni (1985) presented a Lq–Lv–Ls diagram to clarify provenance for cases where the Qm–F–Lt diagram of Dickinson (1985) is not conclusive (Fig. 6c). In this diagram the samples plot mainly in two overlapping fields: 'plate-juncture orogenic highland provenance' (10) and 'continental-arc provenance' (8). Together, the three diagrams are interpreted to indicate provenance from a dissected continental magmatic arc and its basement, partially covered by an epiclastic sedimentary succession. The calculated values of Qp/Q, P/F and V/L ratios (Valloni, 1985) corroborate this interpretation. Smellie (1991) and Andreis, Ribeiro & Trouw (1997) came to a similar conclusion for the Trinity Peninsula Group, Greywacke-Shale Formation and Miers Bluff Formation, and Doktor, Swierczewska & Tokarski (1994) for the Miers Bluff Formation.

5. Metamorphism

Thomson (1973) has given some detailed descriptions of thin sections and also modal analyses from Powell Island, and Dalziel (1984) described the structural, metamorphic and lithological transition between the metamorphic northern part and the non-metamorphic (or much less metamorphic) southern part. However, no detailed description of metamorphism on the island has yet been published. The description given below is based on thin section study and illite crystallinity data.

The lowest grade rocks, at Laurie and Fredriksen islands, show perfectly recognizable detrital grains of quartz, plagioclase, epidote, white mica and biotite. The only visible effect of the metamorphism is the partial decalcification of plagioclase to albite, the incomplete chloritization of clastic biotite, the appearance of tiny metamorphic white mica grains in the matrix and the local occurrence of prehnite, especially in veins and in metavolcanic rocks. In lavas (or subvolcanic rocks), igneous textures are partially preserved, plagioclase grains are albitized and pyroxene (or amphibole) is almost completely substituted by calcite, chlorite and prehnite. The latter generally forms radiate sheaf-like aggregates.

Two metamorphic isograds, marking the appearance of garnet and biotite, were established on Powell Island (Figs 2, 3). The garnet isograd is sharply defined by the sudden appearance of well-developed garnet grains (almandine) in rocks of uniform composition. However, the biotite isograd constitutes a wide zone on the map, covering the range from the first appearance of a few tiny crystals to the general appearance of biotite as a rock forming mineral (marked on the map as 'abundant' biotite). In addition to these isograds, the change in metamorphic grade at Powell Island is also apparent from the gradual increase in both grainsize and quantity of metamorphic white mica. The island can be subdivided by the isograds into four metamorphic zones (Figs 2, 3): zone I without garnet or biotite, comparable to Laurie and Fredriksen islands and also to the Whale Skerries; zone II with garnet but without biotite; zone III with garnet and locally small biotite (<1 %); and zone IV with garnet and relatively abundant biotite (about 5 %).

In zone I, recognizable metamorphic minerals are albite, quartz, chlorite, epidote, white mica, calcite and sphene. In most thin sections, all these minerals occur together, showing that they constitute a stable association. In one thin section from station PO-8, pumpellyite was recognized, in a vein in a metasiltstone. Within zone I a gradual transition takes place from 'sedimentary looking'

Table 2. Total composition of metasandstones from Fredriksen Island, southern Powell Island, the Whale Skerries and Laurie Island (Fig. 2)

	Fredriksen Island										Powell Island			Whale Skerries			Laurie Island								
	FR 1-4	FR 1-17	FR 1-22	FR 2-2	FR 2-5	FR 2-7	FR 2-8	FR 2-11	FR 3-4	FR 3-5	PO 8A-1A	PO 8A-1B	PO 8-1	WS 1-3	WS 1-13	WS 1-17	L 2-3	L 2-9A	L 2-8	L 2-1	L 2-4*	L 3-3	L 3-8	L 2-11	L 1a-1
Qm	36	44	39	37	65	38	39	26	54	59	43	41	40	34	23	3	40	34	39	42	27	44	40	36	28
Qp	6	5	7	7	7	6	6	6	8	6	6	5	13	11	11	10	7	4	3	2	11	8	5	3	5
Or	4	5	4	8	4	4	3	6	2	3	1	1	1	1	-	-	7	8	8	9	5	10	12	13	3
Mc	-	1	1	1	1	2	1	-	1	1	-	-	-	-	-	-	2	2	2	1	-	2	3	4	-
P	22	17	24	29	14	25	21	31	15	15	15	13	9	19	6	10	15	14	30	20	6	21	27	32	12
Lv	9	8	9	6	6	8	8	11	5	5	20	18	10	5	2	2	16	12	9	14	25	6	6	7	23
Ls	-	-	-	-	-	-	-	-	2	2	2	4	7	4	2	5	3	-	-	-	17	-	1	-	18
Lm	3	3	7	4	1	4	3	8	2	2	4	2	3	6	19	8	3	3	5	2	6	1	3	2	10
Micas	1	1	7	6	-	9	14	8	3	4	1	1	1	-	-	-	4	16	5	9	-	1	-	4	-
Ortmtx	19	12	2	2	1	2	4	4	8	3	7	15	16	20	35	29	3	7	-	-	-	7	2	2	1
Pseudm	-	4	-	-	1	1	1	-	-	-	-	-	-	-	-	-	-	-	-	2	-	-	-	-	-

Abbreviations as defined in Dickinson (1985).
* Conglomerate.

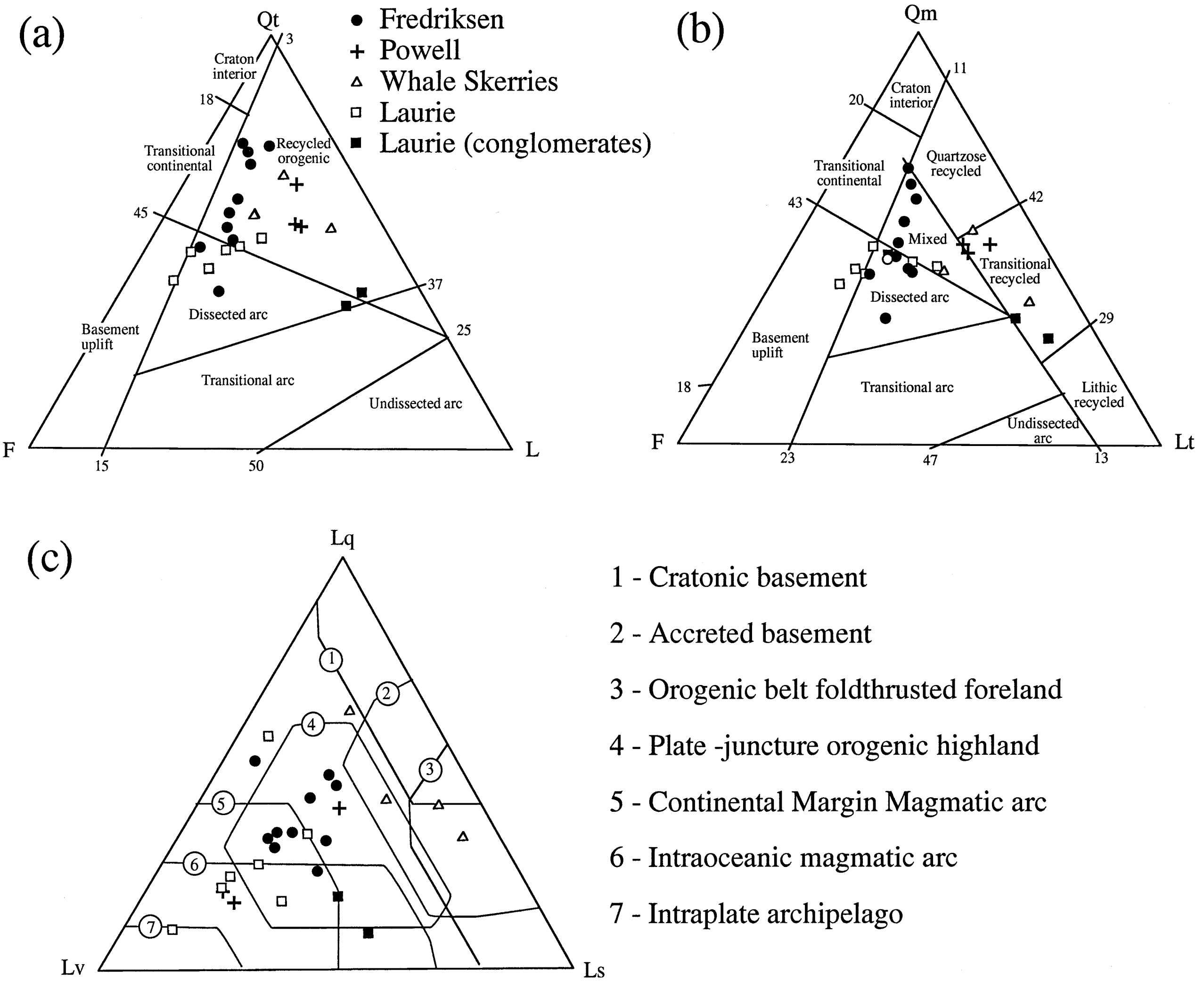

Figure 6. (a) Qt–F–L diagram proposed by Dickinson (1985) to distinguish provenance of arenites, with data of meta-arenites from the South Orkney Islands. (b) Qm–F–Lt diagram proposed by Dickinson (1985) to define the tectonic environment of the source-area of arenites. (c) Lq–Lv–Ls diagram proposed by Valloni (1985) to clarify provenance for cases where the Qm–F–Lt diagram of Dickinson is not conclusive. Lq – quartzose lithics, Lv – volcanic-metavolcanic lithics, Ls – sedimentary–metasedimentary lithics. Further explanation in text.

rocks with recognizable bedding and clast-matrix relationships preserved, to ‘metamorphic looking’ rocks in which secondary foliation and recrystallization predominate over original features. Illite crystallinity data on the islands are published in Trouw, Pankhurst & Ribeiro (1997), including Kubler index values measured on < 2 μm white mica fractions and b_o values measured on rock slabs. The Kubler index values are indicative for temperature (Kisch, 1990) and the b_o values correspond in a broad sense to pressure (Guidotti & Sassi, 1986). They are tentatively plotted on a *PT* diagram (Fig. 7), together with data from the surrounding islands. Although there is some scatter in b_o values, the data are consistent with a gradual increase in metamorphic grade in the following sequence: Laurie and Fredriksen islands – Whale Skerries – zone I of southern Powell Island. Also, a gradual increase in temperature within zone I and between the Greywacke-Shale Formation and the ‘metamorphic complex’ is apparent (Fig. 7).

In zone II the mineralogy of zone I remains stable. The only difference with zone I is that in some thin sections, garnet appears as an additional mineral. In fact there is only one station (PO-7) in this narrow zone.

In zone III the mineral association mentioned for zones I and II remains stable, with the addition of sparce biotite in about half of the examined thin sections. At one station within this zone (PO-6C) stilpnomelane is present as an accessory mineral, commonly associated with calcite. Zone IV is characterized by the presence of biotite as a rock forming mineral in most studied thin sections.

In terms of metamorphic facies the rocks of Powell Island fit in the greenschist facies with a possible gradation in the south to sub-greenschist facies, whereas those from Fredriksen Island can be grouped in the prehnite (pumpellyite) facies. The *P–T* conditions can be estimated as shown in Figure 7, based on a petrogenetic grid (Yardley, 1989; Barker, 1991). Although no typical high pressure minerals have been detected on the island, the

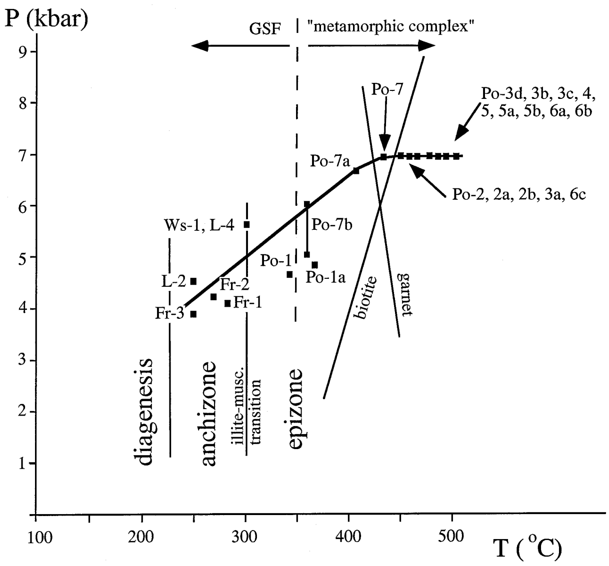

Figure 7. Array of probable peak *P–T* conditions for analysed samples from Powell and surrounding islands; they are based on illite crystallinity data for samples at the lower grade side of the garnet isograd, on Si content in white mica measured by microprobe and as b_o parameters (see text), and on the appearance of garnet before biotite in pelitic assemblages with increasing grade. Stability field of almandine based on Yardley (1989, his fig. 3-11); estimated reaction curve for first appearance of biotite after Nitsch (1970). Note the gradual transition between the Greywacke-Shale Formation (GSF) and 'metamorphic complex' samples.

pressure must nevertheless have been relatively high because of the inversion of the biotite and garnet isograds as compared to standard Barrovian sequences (see, e.g. Yardley, 1989). A similar inversion occurs in the classical Sanbagawa Belt in Japan (Banno, 1986) and has been reported from another part of the Scotia metamorphic complex, at Elephant Island in the South Shetland Islands (Trouw, Ribeiro & Paciullo, 1991). Another indication for moderate to high pressure is the relatively high silica content in white mica from station PO-7B. Eleven microprobe analyses (Table 3) show an Si content per formula unit between 3.31 and 3.39. According to the calibration presented by Massone & Schreyer (1987), later modified by Massone (H. J. Massone, unpub. thesis, Ruhr Univ., Bochum, Germany, 1991), and assuming a metamorphic peak temperature for this zone of about 360 °C, based on the illite crystallinity data discussed above, the pressure would have been between 5 and 6 kbar (Fig. 7).

6. Structures

The main reference structure on Powell Island is a well-developed flat-lying foliation in the Greywacke-Shale Formation, characterized by the approximate parallelism of micas and other platy or acicular minerals, quartz-rich lenses and veins, and lenticular mineral aggregates. In many thin sections, remnants of tight crenulations of an older tectonic foliation can be recognized along the foliation planes. For this reason the main foliation has been labelled S_2, ascribed to a second deformation phase. The orientation of S_2, predominantly dipping weakly to the south, is shown in Figure 8a. A conspicuous stretching lineation, L_2, with a regular NNW–SSE orientation (Fig. 8b) is usually associated with S_2.

In many outcrops S_2 is folded on a centimetric to metric scale by tight to open folds in which generally no new axial planar cleavage is developed. These folds have been labelled D_3. They have axial surfaces that dip moderately to the southwest and axes that plunge

Table 3. Composition of eleven white mica crystals in a sample of grey slate from station PO-7b (Fig. 2)

	1	2	3	4	5	6	7	8	9	10	11
SiO_2	49.10	49.31	48.18	48.65	50.17	48.35	49.91	48.61	50.10	49.70	48.63
TiO_2	0.12	0.13	0.15	0.13	0.07	0.17	0.10	0.13	0.10	0.00	0.12
Al_2O_3	26.93	28.44	27.15	27.32	27.74	27.34	27.15	27.40	27.76	29.82	27.95
Cr_2O_3	0.00	0.00	0.00	0.00	0.00	0.18	0.00	0.00	0.00	0.15	0.00
CaO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
MgO	2.87	2.67	2.77	3.00	3.13	3.33	2.54	2.89	2.62	2.24	2.62
FeO	3.60	4.08	4.49	4.90	4.23	5.20	3.85	4.40	3.16	2.93	3.69
MnO	0.00	0.00	0.00	0.06	0.03	0.00	0.05	0.03	0.00	0.00	0.00
ZnO	0.00	0.34	0.00	0.00	0.37	0.66	0.00	0.00	0.35	0.00	0.00
Na_2O	0.08	0.09	0.11	0.09	0.08	0.12	0.12	0.12	0.12	0.18	0.08
K_2O	10.13	10.26	10.14	9.88	9.84	10.41	10.76	10.30	10.19	10.40	10.14
Total	92.83	95.32	92.99	94.03	95.66	95.76	94.48	93.88	94.40	95.42	93.23
Si	6.74	6.62	6.66	6.66	6.71	6.56	6.78	6.66	6.77	6.63	6.66
Ti	0.01	0.01	0.02	0.01	0.01	0.02	0.01	0.01	0.01	0.00	0.01
Al	4.36	4.50	4.43	4.41	4.37	4.37	4.34	4.43	4.42	4.69	4.51
Cr	0.00	0.00	0.00	0.00	0.00	0.02	0.00	0.00	0.00	0.02	0.00
Ca	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	0.59	0.53	0.57	0.61	0.63	0.68	0.51	0.59	0.53	0.45	0.54
Fe	0.41	0.46	0.52	0.56	0.47	0.59	0.44	0.50	0.38	0.33	0.42
Mn	0.00	0.00	0.00	0.01	0.00	0.00	0.01	0.00	0.00	0.00	0.00
Zn	0.00	0.03	0.00	0.00	0.04	0.07	0.00	0.00	0.03	0.00	0.00
Na	0.02	0.02	0.03	0.02	0.02	0.03	0.03	0.03	0.03	0.05	0.02
K	1.78	1.76	1.80	1.73	1.67	1.81	1.86	1.80	1.75	1.76	1.78

Composition per formula unit is calculated on the basis O = 22.

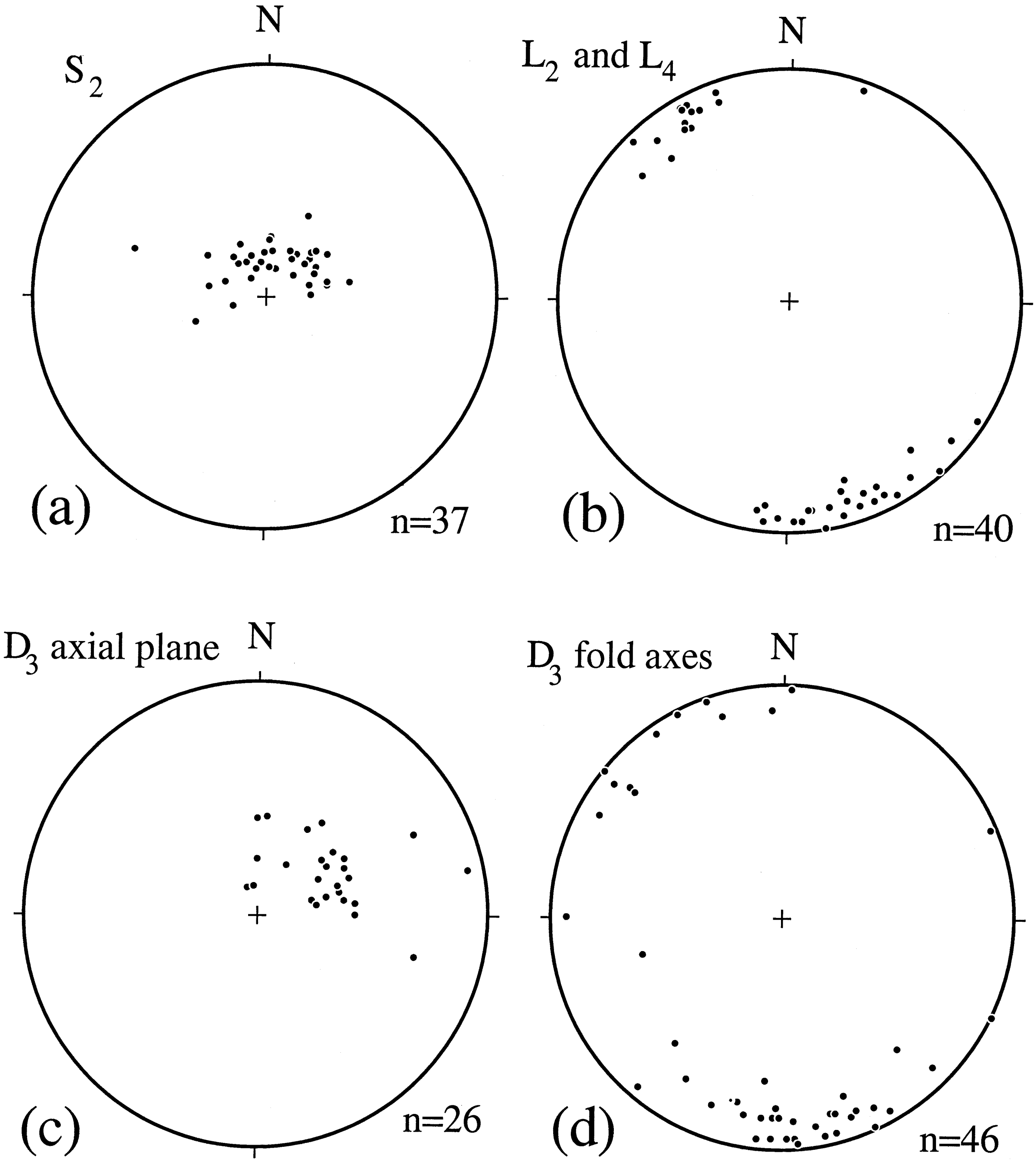

Figure 8. Equal area stereoplots of structures on Powell Island: (a) Poles to S_2 planes; (b) L_2 and L_4 lineations (see text for explanation); (c) poles to axial planes of D_3 folds; (d) D_3 fold axes.

predominantly to the south–southeast (Fig. 8c, d). Many D_3 folds are asymmetric with a vergence to the east.

A conspicuous feature at most outcrops is the presence of shear bands, either developed along the S_2 foliation or, more often, making a small angle with S_2 (Fig. 9). The shear bands are mostly of C´-type (Passchier & Trouw, 1996), but C-type shear bands do also occur. In outcrop and in thin section (Fig. 9), these shear bands define a dominant 'top to the south' sense of shear; of 20 oriented thin sections 17 gave this result, two gave the opposite result and one gave shear in both senses. The presence of conflicting shear sense markers may indicate that non-coaxial flow locally deviated from simple shear, for example as a 'stretching shear zone' (Means, 1989; Passchier, 1991).

The shear bands indicate a down-dip normal movement, attributed to D_4. The stretching lineations associated with this phase (L_4), interpreted to reflect the main movement direction, are subparallel to L_2 lineations (Fig. 8b). In the field they are virtually indistinguishable but in thin section the difference is apparent from the smaller grain size of recrystallized quartz in L_4 lineations, probably reflecting lower temperature conditions of deformation (Passchier & Trouw, 1996).

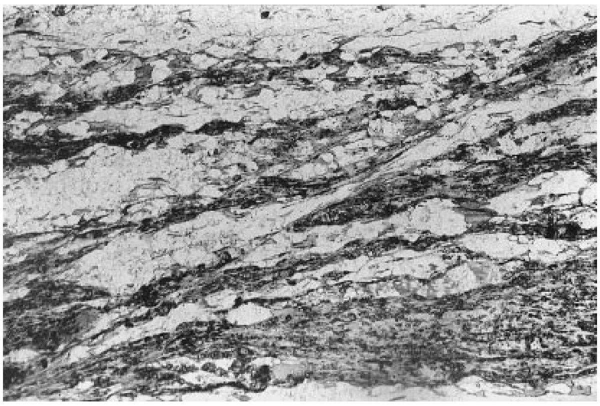

Figure 9. Photomicrograph of D_4 shear bands in phyllite, indicating sinistral sense of shear. Station PO-2. Width of view 6 mm. Plain polarized light.

D_5 is a phase of brittle faults and kink bands that are usually steeply dipping and which predominantly have east–west or north–south trend. They indicate mostly horizontal extension, although some constrictional faults have been observed as well. Station PO-1 is the link between the low grade southern part of Powell Island, where S_0 is well preserved, and the northern part where S_0 is transposed. At this station, S_0 and structures of all five deformation phases can be observed.

The structures on Powell Island were described by Dalziel (1984) as belonging to three deformational phases: an early phase, a main phase and a late phase. His early phase correponds to our D_1, his main phase to our D_2 and his late phase to our combined D_3, D_4 and D_5.

Meneilly & Storey (1986) reported a similar sequence of five deformation phases from Signy Island, interpreted by them to reflect tectonic transport of the hanging wall to the north or north–northwest, related to subduction in the opposite direction. No indication of normal movement with down-dip tectonic transport to the south is described by them; our observations from two outcrops on the south coast of Coronation Island, northeast of Signy Island (Fig. 1b), confirm the lack of such structures at these sites. It therefore seems that the south-directed, normal movements related to D_4 are restricted to Powell Island. However, it is possible that such structures exist elsewhere on Coronation Island, which has not been investigated in detail.

In the lower grade rocks of southern Powell Island, the Whale Skerries and Fredriksen Island the structures are less regular and will therefore be described separately below. The main structure at Fredriksen Island is a penetrative cleavage, S_1, (sub)parallel to the well-preserved bedding S_0, dipping predominantly to the west (Figs 2, 4). Few isoclinal small scale (< 1 m) D_1 folds with axes 290/10 show folded S_0 with S_1 along the axial surface. As shown in Figure 4, S_0/S_1 are folded in open to gentle folds with steep axial planes (250/90) and axes 354/10.

At the Whale Skerries, S_0 can also be clearly recognized with an S_1 cleavage at a small angle or subparallel to it.

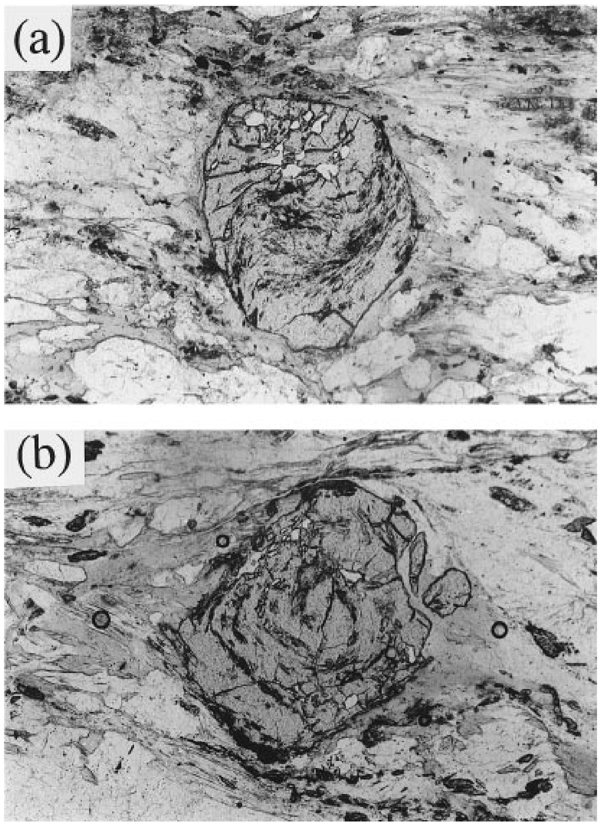

Figure 10. (a) Photomicrograph showing garnet crystal with an inclusion pattern interpreted as helicitic D_2 folds; although S_i is discontinuous with S_e the garnet is interpreted to have grown during D_2, since the S_2 foliation in the matrix (S_e) is deflected around the garnet. Station PO-5. Width of view 3 mm. Plain polarized light. (b) Garnet crystal with a spiral inclusion pattern suggesting syntectonic growth; this may refer either to D_1 or to D_2. Station PO-6a. Width of view 3 mm. Plain polarized light.

Some isoclinal small scale D_1 folds have northwest–southeast trending axes. S_0 and S_1 are folded in large scale (> 10 m) open folds with northwest–southeast axes and steep standing axial surfaces. A large number of steep faults with fault breccias and variable orientation result in a somewhat chaotic aspect of many outcrops.

7. Comparison of structural evolution and metamorphism

The relative growth period of metamorphic minerals with respect to the deformation phases is as follows. S_2 surfaces show a strong deflection around garnet porphyroblasts (Fig. 10), indicating that these grew before or during D_2. Inclusion patterns in garnet include apparent folds of an older foliation (S_1; Fig. 10a) and spiral shapes (Fig. 10b). The first pattern suggests garnet growth contemporaneous with D_2, but the second may be explained both as syn-D_1 and syn-D_2 garnet growth. It can be concluded that garnet growth may have started during D_1, was definitely in progress during D_2, and stopped before the end of this phase (Fig. 11). The occurrence of epidote, calcite, albite and sphene inclusions within the

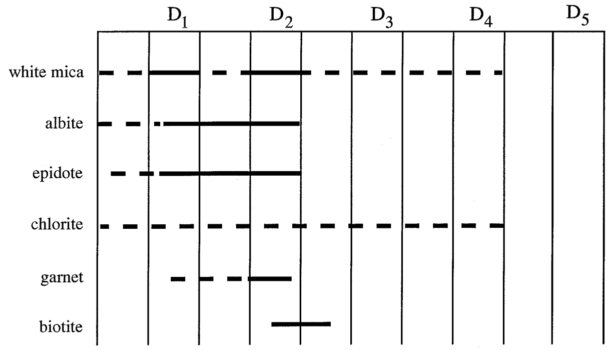

Figure 11. Growth periods of the main metamorphic minerals with respect to the deformation phases.

garnets demonstrates that the formation of these minerals preceded garnet growth, although partial contemporaneity cannot be excluded. The growth period of biotite is more difficult to interpret because this mineral did not form porphyroblasts nor does it occur as inclusions. However, local substitution textures suggest that biotite grew later than garnet. Figure 11 summarizes the relations and shows that peak metamorphic conditions were attained during D_2, with deformation under retrograde conditions during the later phases.

8. Age data

Previously published age data from the Scotia metamorphic complex of the South Orkney Islands (all from Signy and Coronation islands) fall mainly in the range 176–205 Ma. Miller (1960) obtained K–Ar ages of 176 Ma in muscovite and between 176 and 199 Ma in biotite. Grikurov, Krylov & Silin (1967) determined an age of 205 Ma for muscovite by the same method, and Tanner, Pankhurst & Hyden (1982) published an age of around 190 Ma for a hornblende separate, also determined by the K–Ar method. Ar–Ar analysis in muscovite yielded an age of 184 ± 4 Ma (Grunow *et al.* 1992). However, a Rb–Sr ‘errorchron’ indicates a much older age of 281 ± 56 Ma (Tanner, Pankhurst & Hyden, 1982; Rex, 1976).

The supposed Permo-Triassic age of the Greywacke-Shale Formation, based on its correlation with the Trinity Peninsula Group (Thomson, 1975) is supported by Triassic radiolarians reported by Dalziel *et al.* (1981) from an isolated chert occurrence at Scapa Rock (Fig. 1b), interpreted as part of the Greywacke-Shale Formation. Rb–Sr isochron ages were reported by Pankhurst (1983; 281± 16 Ma) for the Trinity Peninsula Group at Hope Bay and by Willan, Pankhurst & Hervé (1994; 243 ± 8 Ma) for the supposedly correlative Miers Bluff Formation at Livingston Island, South Shetland Islands.

Trouw, Pankhurst & Ribeiro (1997) obtained an ‘errorchron’ of 213 ± 38 Ma with MSWD of 12 from seven samples of deformed phyllites from Powell Island (station PO-7B; Fig. 2). A new whole rock K–Ar age of 239 ± 19 Ma is reported here from an anchizone metamorphosed volcanic pebble in the diamictite at Fredriksen Island (FR-1-8A). Although this age conforms well with other ages obtained in this area, the potassium content of this pebble is relatively low (0.0791 %) and atmospheric Argon content is high (Ar_{atm} 76.25 %).

In the context of the geological relations outlined above, available radiometric and fossil ages are here interpreted as follows:

The sedimentation age of the Greywacke-Shale Formation is probably Triassic as indicated by the fossils from Scapa Rock (Dalziel *et al.* 1981), fossils from the Cape Legoupil area (Thomson, 1975) and by most of the Rb–Sr isochrons from Greywacke-Shale Formation and equivalent units (Trinity Peninsula Group and Miers Bluff Formation). If the K–Ar age (239 ± 19 Ma) of the volcanic pebble mentioned above is interpreted as its crystallization age, the sedimentation of the Greywacke-Shale Formation must have occurred between about 200 (age of metamorphism) and 239 ± 19 Ma. However, if this age is understood as representing metamorphism (improbable, because of the other metamorphic ages mentioned above), or a mixed value between metamorphism and crystallization, the sedimentation might have initiated before 239 Ma. The metamorphism of the Greywacke-Shale Formation and ocean floor material (see Section 9.a) at Signy and Coronation Islands probably occurred in the interval 176–205 Ma, as indicated by all K–Ar mineral ages and by the Ar–Ar data. The Rb–Sr ‘errorchron’ from Powell Island (213 ± 38 Ma) with its large error could either reflect sedimentation or metamorphism or even represent a mixture of both. Trouw, Pankhurst & Ribeiro (1997) interpreted it as a sedimentation age. By the late Jurassic (about 150 Ma) the metamorphosed sequences had been uplifted and were cropping out and eroding as testified by remnants of alluvial fans from this period with metamorphic detritus (Powell Island Conglomerate; Elliot & Wells, 1982; Wells, 1984).

The Rb–Sr ‘errorchron’ of 281 ± 56 Ma from Coronation and Signy islands (Tanner, Pankhurst & Hyden, 1982; Rex, 1976) and the similar isochron of 281 ± 16 (Pankhurst, 1983) from the Trinity Peninsula Group at Hope Bay, do not fit in the scheme presented above, and are difficult to explain. Similar values were reported from the Elephant Island group (Hervé *et al.* 1990, 1991; Trouw, Pankhurst & Kawashita, 1990), South Shetland Islands, which is also part of the Scotia metamorphic complex; these were interpreted by Trouw, Pankhurst & Kawashita (1990) as possibly due to the presence of relic radiogenic Sr from the source area.

9. Discussion

9.a. Comparison with Coronation and Signy islands

The interpretation of the Scotia metamorphic complex at northern Powell Island as a metamorphosed part of the Permo-Triassic Greywacke-Shale Formation raises the question as to what extent this interpretation is also valid for other parts of the Scotia metamorphic complex (Dalziel, 1982, 1984). Within the South Orkney Islands, the metamorphic sequence of northern Powell Island is certainly quite different from the sequence that crops out on Signy Island (Storey & Meneilly, 1985) (Fig. 1b). Whereas the former is principally composed of continent-derived turbidites, the latter contains a considerable percentage of metachert, marble and mafic rocks of ocean floor origin (Storey & Meneilly, 1985). The data available from Coronation Island (Thomson, 1974) (Fig. 1b) show at least in part a close similarity to the ones from Signy Island. During our investigations four stations were visited on Coronation Island (Fig. 1b). Preliminary results suggest the existence of a gradual transition between predominantly ocean floor-derived sequences in the central part, to continent-derived submarine fan successions in the eastern part of the island. Hence, in the South Orkney Islands the equivalence between the 'Scotia metamorphic complex' and the Greywacke-Shale Formation is only valid for Powell Island and possibly for a small segment of eastern Coronation Island.

9.b. Comparison with the South Shetland Islands

Other parts of the Scotia metamorphic complex that crop out in the South Shetland Islands (Fig. 1a) have been interpreted as being, at least in part, derived from ocean floor material (Dalziel, 1984; Trouw, Pankhurst & Kawashita, 1990; Grunow *et al.* 1992; Valeriano & Heilbron, 1994). These parts have a similar composition and tectonic setting to the rocks on Signy and Coronation islands, but they yield different radiometric ages. Based on these ages a threefold subdivision of the complex has been proposed (Trouw, 1991; Trouw *et al.* 1994): the South Orkney Islands (180–220 Ma) and, in the South Shetland Islands, the Elephant Island group (80–120 Ma) and Smith Island (47–53 Ma).

The metamorphic sequence from Powell Island can only be correlated with metamorphic sequences from the South Shetland Islands in terms of a similar tectonic setting, related to the Pacific subduction zone of Gondwana. However, there are significant differences in protolith and in age of metamorphism. The Scotia metamorphic complex is a tectono-metamorphic unit rather than a litho-stratigraphic one and it may therefore include rocks of different ages and protoliths. However, the Greywacke-Shale Formation and its higher metamorphic equivalent on Powell Island are still comparable to the Trinity Peninsula Group on the Antarctic Peninsula, and to the Miers Bluff Formation in the South Shetland Islands.

9.c. Mesozoic evolution of the South Orkney Microcontinent

The Mesozoic evolution of the SOM has been regarded as relatively straightforward by previous authors. Dalziel (1984) described north–south trending lineations in the South Orkney Islands and attributed them to a subduction-related tectonic regime in the early Mesozoic. King & Barker (1988) described a major graben south of the islands filled with 5 km of sediments, the Newton graben, as a basin associated with a fore-arc setting. Barker, Dalziel & Storey (1991) interpreted the east–west trending structure of the islands, the presence of the Newton graben and an east–west trending magnetic anomaly south of this graben (Harrington, Barker & Griffiths, 1972) as an accretionary prism, fore-arc spreading basin and magmatic arc triplet associated with south-directed subduction (in the present orientation). Our data show that the situation may be more complicated than suggested by any of these authors. Our D_1–D_3 structures on Powell Island, similar structures on Coronation Island and possibly the foliations on Fredriksen and Laurie islands are thought to be associated with south-directed (according to present coordinates) subduction in early to middle Jurassic times. Evidence for such south-directed subduction has also been found on Signy Island (Meneilly & Storey, 1986). The close association of these structures with the relatively high pressure metamorphism suits this interpretation well. The dominant east-vergent asymmetry of D_3 folds on Powell Island may be due to a gradual change in subduction direction during which older planar structures rotated into the shortening field, while stretching lineations kept developing without being refolded.

Our D_4 structures have a similar direction but opposite shear sense to the subduction movement and are apparently associated with horizontal extension and uplift; D_4 is a phase of ductile mylonitization at lower greenschist facies conditions and can be dated as middle to late Jurassic. By late Jurassic time, the entire sequence had been uplifted to the surface and was covered unconformably by terrestrial sediments (Thomson, 1981), as elsewhere throughout the Scotia arc region (Dalziel, 1984). Consequently, D_4 is interpreted as an important phase of north–south crustal extension. The geometry of D_4 shear bands suggests that the zone in which D_4 structures formed was a stretching shear zone (Means, 1989), and such shear zones may be typical for large-scale crustal extension (Passchier, 1991).

D_4 structures have so far only been found on Powell Island. Coronation, Powell, Fredriksen and Laurie islands represent blocks of different metamorphic grade and structural evolution. These differences may repesent various structural regimes that were present at different depths in the crust before uplift. During peak metamorphic conditions, Coronation Island may have been at a deeper crustal level than Powell Island, and Powell at a deeper level than any of the eastern islands (Fredriksen, Laurie and smaller islands). The fact that D_4 ductile extensional structures are limited to Powell Island may indicate that they developed in a shear zone of limited width at the crustal level of Powell Island, above the level of Coronation Island and below that of Fredriksen, Laurie and the eastern islands. The differences in metamorphic grade between the islands may partly be due to juxtaposition during the extensional D_4 event.

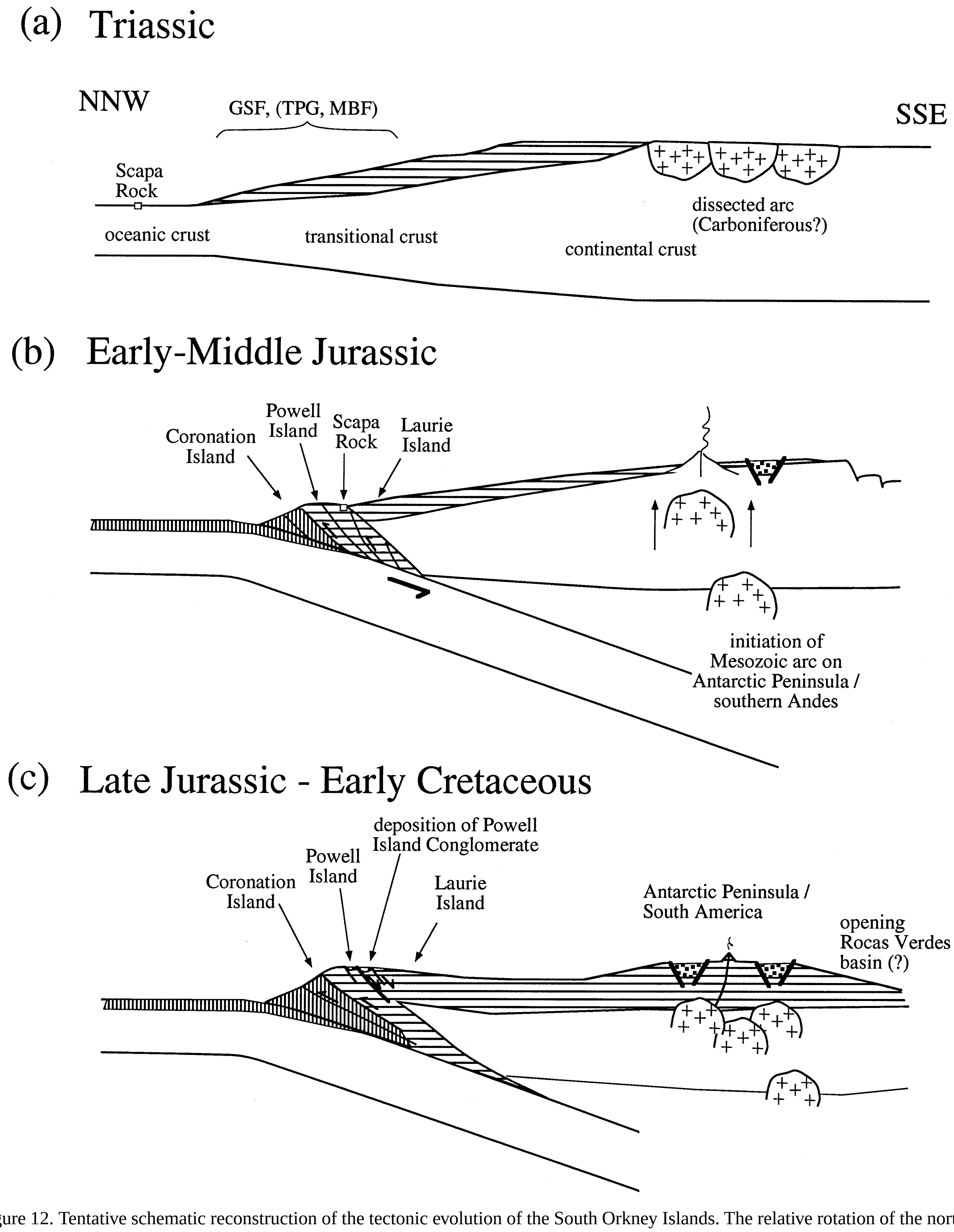

Figure 12. Tentative schematic reconstruction of the tectonic evolution of the South Orkney Islands. The relative rotation of the northern part of the Antarctic Peninsula and of the South Orkney Islands block with relation to Powell Basin from which it was torn off is not taken into consideration in this reconstruction. (a) Deposition of Greywacke-Shale Formation (GSF), Trinity Peninsula Group (TPG) and Miers Bluff Formation (MBF) in Triassic times along an inactive continental margin with a dissected arc in the source area. (b) Deformation and metamorphism of the rocks underlying Powell Island, related to the development of a subduction zone, during early to middle Jurassic times. The sequences of Coronation Island, mainly derived from ocean floor material (vertical striping), suffered metamorphism at the same time. (c) During late Jurassic–early Cretaceous times the rocks of Powell Island suffered extensional deformation followed by uplift and erosion. The Powell Island Conglomerate was deposited as alluvial fans.

In their present orientation, the D_4 structures on Powell Island are exposed on topographic levels where they should continue into the neighbouring islands. The fact that this is not the case shows that differential uplift along the major north–south trending extensional brittle faults, between and on the islands (Fig. 2), must also have played a (possibly minor) role in the present juxtaposition of the islands. These faults cut the conglomerates and may be associated with the separation of the SOM and the Antarctic Peninsula during development of the Scotia arc. The Newton graben has been interpreted as a fore-arc basin (King & Barker, 1988), but could also be a D_4 structure.

9.d. Summary of evolution

The tectonic evolution of the South Orkney Islands can now be summarized as follows (Fig. 12). In (Permo-)Triassic times, submarine fans accumulated along a probably inactive continental margin (Smellie, 1991; Doktor, Swierczewska & Tokarski, 1994; Andreis, Ribeiro & Trouw, 1997) next to a continent with a dissected magmatic arc (Fig. 12a). As discussed by other authors (e.g. Arche, Lopez Martínez & Marfil, 1992; Smellie, 1991; Birkenmajer, 1992), this continent may have included southern South America. Subsequent subduction (early–middle Jurassic, 180–200 Ma) led to the metamorphism of part of the ocean floor and its cover (Scotia metamorphic complex) including part of the turbidites (Powell Island), accompanied by deformation (D_1, D_2 and probably D_3; Fig. 12b). Later extension (middle–late Jurassic) in the north–south direction with respect to the present position of the islands caused D_4 structures at Powell Island under retrograde metamorphic conditions related to uplift (Fig. 12c). These movements resulted by the end of Jurassic time in the exposure of the metamorphic sequence, as testified by remnants of alluvial fan deposits (Spence Harbour and Powell Island Conglomerate; Elliot & Wells, 1982; Wells, 1984) of late Jurassic to early Cretaceous age (Fig. 12c). The final uplift and erosion in Cenozoic times was accompanied by brittle deformation (D_5) that affected the conglomerates as well.

9.e. Regional geological implications

Recent reconstructions of the southern Scotia arc have shown that relative movement between southern South America and the Antarctic Peninsula, eventually leading to separation of the SOM from adjacent continental fragments, started at about 100 Ma (Grunow *et al.* 1992; Cunningham *et al.* 1995). With the possible exception of D_5, the structures and metamorphism on Powell and neighbouring islands all pre-date deposition of the Powell Island Conglomerate in late Jurassic time, and therefore pre-date development of the Scotia arc. The significance of these early fabrics can be better assessed if later evolution of the SOM during development of the Scotia arc is considered. For this purpose, it is necessary to consider both the structures on the islands, and those on and around the submerged parts of the SOM.

The SOM has undergone considerable extension and is composed of an unusually thin crust (King & Barker, 1988). King & Barker (1988), Rodriguez-Fernandez *et al.* (1994) and King *et al.* (1994) investigated the structure of the submerged part of the SOM and its margins. They found dominant north–south trending faults and small sedimentary basins throughout the SOM, which represent east–west extension. These structures post-date east–west trending faults and the major east–west trending Newton graben south of the islands, which represents north–south extension.

The Powell Basin (Fig. 1a) has been presented as a small extensional oceanic basin that formed by east or northeast directed movement of the SOM away from the Antarctic Peninsula between 40 and 23 Ma (King & Barker, 1988; King *et al.* 1994; Lawver, Williams & Sloan, 1994). Strike-slip faults seem to bound the northwest side of the Powell Basin towards the South Scotia Ridge, while the southwest margin of the basin towards the Antarctic Peninsula appears to be of a transtensional nature (Fig. 1a; Rodriguez-Fernandez *et al.* 1994; Galindo-Zaldivar *et al.* 1994). The South Scotia Ridge is deformed by sinistral strike-slip faults that connect small extensional basins (Galindo-Zaldivar *et al.* 1994). Although no clear magnetic anomalies have been found on the oceanic floor of the Powell Basin, there is some indication of a northwest–southeast oriented spreading ridge (McAdoo & Marks, 1992).

From the available offshore data it seems that the SOM was attached to the tip of the Antarctic Peninsula until approximately 40 Ma, and it then drifted eastward in response to major extension in the Powell Basin and in

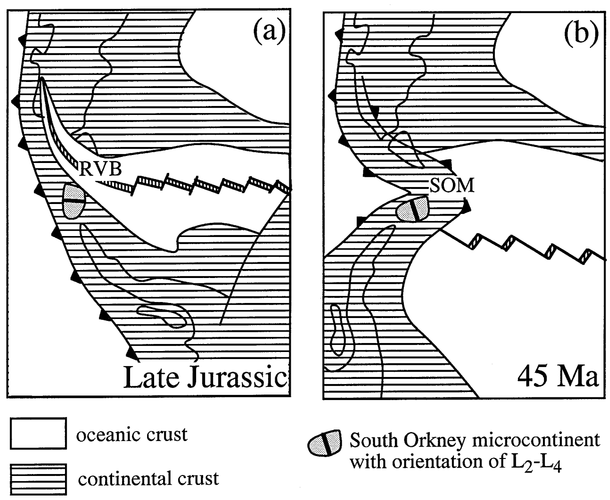

Figure 13. Schematic reconstruction of the Scotia arc region modified after Barker, Dalziel & Storey (1991). (a) Late Jurassic; (b) 45 Ma. Symbols as in Figure 1a. Explanation in text. RVB – Rocas Verdes Basin; SOM – South Orkney Microcontinent.

the microcontinent. Major north–south trending D_5 extensional faults between and on the islands are probably of this age since they cut the conglomerates. The reconstruction of movement of the SOM during development of the Scotia arc indicates that it probably did not rotate much during its separation from the Antarctic Peninsula, but that it moved to the east or northeast along the South Scotia transform fault and another transcurrent fault bounding the SOM to the south. This implies that the north–south or northwest–southeast trending lineations on Powell Island and the gently plunging foliations were in approximately their present orientation at the onset of Scotia arc development.

9.f. Implications related to the breakup of Gondwanaland

The sequence of early constrictional and late extensional movements in a north–south direction on Powell Island can hardly be dismissed as a coincidence, especially since they are separated by a relatively limited time interval. D_4 extension may be an effect of gravity collapse in a developing accretionary wedge (Fig. 12). Another possibility is that D_4 is an effect of the first stages of breakup of Gondwanaland that occured around this time (middle to late Jurassic). Reconstructions of the first opening between east and west Gondwana suggest that spreading did not interrupt the Pacific active continental margin that connected South America and the Antarctic Peninsula at the time, but that this spreading led to the formation of the Rocas Verdes Basin in southern South America (Fig. 13a) (Barker, Dalziel & Storey, 1991). The extension observed in the SOM may therefore alternatively be associated with opening of the Rocas Verdes Basin (Fig. 13a). In late Jurassic time, the Antarctic Pensinsula and the still attached SOM have been reconstructed to lie approximately parallel to the coast of South America (Fig. 13a) (Barker, Dalziel & Storey, 1991). In this orientation, the SOM would have been in a position somewhere between the Pacific continental margin and the proposed position of the Rocas Verdes Basin, and L_2–L_4 lineations would have been in a near east–west orientation, slightly oblique to both (Fig. 13a) (Barker, Dalziel & Storey, 1991). D_4 extension could have been directly related to the extension that created the Rocas Verdes basin on the internal part of the Pacific active margin.

King & Barker (1988) and Barker, Dalziel & Storey (1991) suggest that clockwise rotation of the Antarctic Peninsula with respect to South America in the latest Cretaceous, due to opening of the Weddell sea and the ocean between West and East Gondwana, caused north–south convergence in the Pacific margin of Gondwana to form a large-scale east-directed cusp-fold structure (Fig. 13b). The SOM would be located on the southern limb of this 'pre-Scotia cusp', opposite South Georgia. This rotation would have brought the SOM roughly into its present orientation, which it must have had around 40 Ma (Fig. 13b; King & Barker, 1988). Although the model sketched above is attractive, we found no structures that we could associate with this rotation of the SOM. If the rotation took place as envisaged by King & Barker (1988) and Barker, Dalziel & Storey (1991), deformation must have taken place along shear zones which happen to lie outside the presently exposed part of the SOM. The alternative model presented by Cunningham *et al.* (1995) seems to be in better agreement with the observed structures.

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