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MAP COMPILED BY J.W. THOMSON
TEXT BY D.I.M. MACDONALD and B.C. STOREY

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GEOLOGICAL MAP OF SOUTH GEORGIA
SUPPLEMENTARY TEXT

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Abstract

The island of South Georgia is the emergent part of an isolated microcontinental block. Its geology can be matched closely to that of western Patagonia and Tierra del Fuego, the South Georgia block apparently having formed part of the Pacific margin of Gondwana. For much of the Mesozoic the south-western edge of Gondwana was an active plate margin above the easterly subducting, proto-Pacific ocean floor. The rocks of South Georgia represent various stages in the evolution of the active margin, in particular the development of an island-arc–back-arc-basin system of middle Jurassic to mid-Cretaceous age.

The oldest rocks are polyphase-deformed metasediments which probably formed in a pre-Middle Jurassic accretionary complex. They were intruded by an extensive tholeiitic suite of gabbroic plutons which migmatized the country rock. This magmatic episode thinned the continental crust; further rifting and emplacement of tholeiitic magma led to the creation of mafic crust (composed of lavas, dykes and plutons) with oceanic crustal characteristics. The middle–late Jurassic magmatic episode is represented by an igneous and metamorphic complex and an ophiolite suite, now exposed in the southern part of South Georgia. It is likely that a similar mixture of thinned continental crust and oceanic crust forms much of the hidden floor of the marginal basin.

During the early Cretaceous an island-arc–back-arc-basin system was active in the area. The island-arc suite, a series of mudstones with interbedded tuffs and overlying volcaniclastic breccia, crops out on isolated islands off the south-west coast. All of the pyroclastic rocks are andesitic and have the calc-alkaline chemistry typical of island arcs. Deposition of the clastic rocks began in the latest Jurassic and continued into the late Aptian or Albian; intrusions of dioritic stocks, sills and dykes of late Cretaceous age represent a late phase of igneous activity.

The age of deposition in the marginal basin is poorly constrained but probably contemporaneous with deposition of the arc sediments. Most of South Georgia is formed of a *c.* 6 km thick unit of Lower Cretaceous andesitic greywackes, derived from the volcanic arc and deposited by turbidity currents. These flowed both across and along the basin, forming a laterally variable complex which was penetratively deformed early in the late Cretaceous. Deformation was associated with closure of the basin and westerly underthrusting of the basin floor. The volcaniclastic turbidites were thrust over a series of siliciclastic turbidites of unknown age which may have been derived from the opposite (continental) side of the basin.

The arc and the basin were separated by a major fault during deposition. The line of this is now marked by a mylonite zone which records both dip-slip and strike-slip movement. This is the only unit in the island younger than mid-Cretaceous.

Marine geophysical data from the continental block indicate the continuity of terranes offshore and reveal the presence of a large magnetic batholith on the south-western fringe of the block. This has very similar characteristics to the Patagonian batholith and largely confirms the former presence of South Georgia on the Pacific side of the Andean cordillera. Displacement from the southern margin of the South American continent to its present isolated position was caused by formation of the Scotia Sea during Tertiary times.

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PREFACE

This memoir is a review of current knowledge of all the geological formations of South Georgia, discussed in their regional context and presented within the framework of the geological evolution of the area. Both the geological map and its accompanying text are based on original field work by a large number of geologists (Table I), mostly members of the British Antarctic Survey (BAS). The main programme of BAS field work lasted from 1969 to 1977, with a change of emphasis from regional mapping to specialist studies after 1973. Laboratory work continued until 1980, and most of the major findings were published by 1986. The South Georgia project was the first detailed thematic study undertaken by BAS, and publication of this map and memoir marks the end of a distinct phase in the geological investigation of South Georgia.

The memoir is dedicated to Alec Trendall, who showed us all the way.

Table I. Summary of geological investigations on South Georgia since 1951

Name	*Date*	*Area*	*Comments*
A.F. Trendall	Nov.–Dec. 1951	Cumberland Bay & Royal Bay; parts of interior; circumnavigation of island	South Georgia Survey 1951–2
A.F. Trendall	Oct. 1953–April 1954	North coast; landings along west coast; south-east part of island	South Georgia Survey 1953–4
N. Aitkenhead & P.H.H. Nelson	Dec. 1959–Jan. 1960	Cumberland West Bay–Cape George	British Antarctic Survey (BAS)
M.J. Skidmore	Nov. 1966–Jan. 1967	Stromness Bay–Fortuna Bay	BAS
M.J. Skidmore	Feb.–April 1969	Prince Olav Harbour area	BAS
P. Stone	Dec. 1970–March 1971	Dartmouth Point–Barff Point–Hound Bay	BAS
P. Stone	Nov. 1971–March 1972	St. Andrews Bay–Gold Harbour	BAS
R.B. Crews	Dec. 1971–April 1972	Royal Bay & Cumberland East Bay	BAS
R.A.S. Clayton	Nov. 1972–April 1973	Bay of Isles	BAS
T.H. Pettigrew	Dec. 1972–April 1973	Annenkov Island	BAS
I.W.D. Dalziel, R.L. Bruhn, R.H. Dott Jr. & R.D. Winn Jr.	Jan.–March 1973	Cumberland East Bay–Stromness Bay	US Antarctic Research Program
P. Stone	Nov. 1973–March 1974	St. Andrews Bay–Cooper Bay	BAS
R.A.S. Clayton	Dec. 1973–April 1974	Ice Fjord–King Haakon Bay	BAS
R.N. Mortimore	Dec. 1973–April 1974	Ice Fjord–King Haakon Bay	BAS
P.W.G. Tanner	Nov. 1973–Feb. 1974	Shag Rocks; Elsehul; Prince Olav Harbour–Stromness Bay; Clerke Rocks; landings on southern and south-western South Georgia	BAS
C.M. Bell, B.F. Mair & B.C. Storey	Nov. 1974–April 1975	Southern South Georgia	BAS
B.C. Storey	Nov. 1975–April 1976	Drygalski Fjord–Cape Vahsel; parts of interior; islands off west coast; Ducloz Head; Diaz Cove	BAS
P.W.G. Tanner	Nov. 1975–April 1976	Fortuna Bay–Cumberland East Bay; islands off west coast	BAS
D.I.M. Macdonald	Nov. 1975–April 1976	Cape Darnley–Queen Maud Bay; islands off west coast	BAS
D.I.M. Macdonald	Oct. 1976–April 1977	Queen Maud Bay; northern South Georgia; parts of interior; Annenkov Island	BAS
B.F. Mair	Nov. 1976–April 1977	Larsen Harbour–Leon Head; Dartmouth Point area; Annenkov Island	BAS
D. Craw & I.M. Turnbull	Nov.–Dec. 1984	Ross, Hindle & Weddell glaciers; Royal Bay	NZ South Georgia Expedition
I.W.D. Dalziel, D.G. Beaver, H.K. Brueckner, A.M. Grunow, A.W. Meneilly & S.B. Mukasa	June 1985	Landings on southern South Georgia	US Antarctic Research Program

GLOSSARY OF ABBREVIATIONS

Drygalski Fjord Complex	DFC
Salomon Glacier Formation	SG
Cooper Island Formation	CI
Novosilski Glacier Formation	NG
Larsen Harbour Complex	LHC
Larsen Harbour Formation	LH_1
Annenkov Island Formation	AF
Lower Tuff Member	AF_1
Upper Breccia Member	AF_2
Cumberland Bay Formation	CB
Sandebugten Formation	SF
Cooper Bay Formation	CO
Ducloz Head Formation	DH
Coastal Member	DH_2
Inland Member	DH_1
Cooper Bay dislocation zone	CBDZ
Shackleton Fracture Zone	SFZ

1 Introduction

South Georgia is a mountainous island which lies 2000 km due east of Cape Horn, within the Antarctic Convergence (Fig. 1). The island is *c.* 170 km long and varies from 2–40 km wide. The highest peak is Mount Paget (2934 m) and there are many mountains of about 2000 m or more. The isolation and rugged nature of South Georgia ensured that exploration and geological mapping progressed very slowly after the discovery of the island by Captain James Cook in 1775.

No geological observations were made until 1882, when the German International Polar Year Expedition wintered in Royal Bay (Will, 1884). The island was investigated by a number of geologists between 1882 and the Second World War; a complete record of this early work is given by Trendall (1953). Unfortunately most of the early expeditions only had a peripheral interest in South Georgia and little of the work was systematic. The main rock types were noted early on (Thurach, 1890; Nordenskjöld, 1905; Heim, 1912) and it was quickly realized that the island is dominantly formed of metamorphosed sediments of various types, with igneous rocks in the south and west. The presence of large-scale folding on axes parallel to the length of the island was noticed (Andersson, 1907) and analogies were made with the geology of the southern Andes (Nordenskjöld, 1905). A few fossils were also found, and there was much debate about the age of the sedimentary rocks. Estimates ranged from Ordovician (Gregory, 1914) to Cretaceous (Wilckens, 1937, 1947). Attempts were made to construct stratigraphic schemes (Ferguson, 1915; Wordie, 1921; Holtedahl, 1929; Tyrrell, 1930) but many of these efforts were unsuccessful due to a lack of fossils and the patchy nature of much of the sampling. Undoubtedly the best of the early work was done by Tyrrell who first described material collected by Ferguson. This stimulated a long interest in the petrography of the Scotia arc, including six papers on South Georgia (Tyrrell, 1914, 1915, 1916, 1918, 1930, 1945). His petrographic descriptions were unrivalled, and form the basis for one of the best of the early stratigraphic schemes (Tyrrell, 1930), even though he never visited the island.

The first detailed geological investigations were made by Trendall (1953, 1959) who accompanied the South Georgia Survey Expeditions on two occasions during their four-season survey (1951–57). Despite having to accommodate to the needs of a surveying party he covered a great deal of ground in some detail, and was the first geologist to undertake structural mapping and sedimentological studies.

A British Antarctic Survey programme to map the whole island started in 1969 and involved eleven geologists over a period of eight years, resulting in many publications describing aspects of local areas (e.g. Skidmore, 1972; Stone, 1980; Pettigrew, 1981). Skidmore (1972) gave details of work undertaken on the island in the 1950s and 1960s. A party from Lamont-Doherty Geological Observatory of Columbia University visited South Georgia during the 1973–74 season and did sedimentological and structural work on the north-eastern coast (Dott, 1974; Dalziel and others, 1975; Winn, 1978). Drawing on their own work, and on work by BAS geologists (notably T.H. Pettigrew), they made comparisons with the geology of southern South America and suggested that all the rock groups present on South Georgia have counterparts in Tierra del Fuego, and that all can be referred to parts of the Andean orogenic cycle. This has proved to be a very fruitful model and it has been largely confirmed by the later phases of BAS mapping (Bell and others, 1977; Storey and others, 1977; Tanner and others, 1981) and age dating (Tanner and Rex, 1979; Thomson and others, 1982).

As well as land-based geology there have been some marine investigations. The first of these was the dredging programme of the "Discovery" expeditions (Tyrrell, 1945). More recently, workers from Birmingham University have carried out extensive marine geophysical investigations in the Scotia Sea and South Atlantic (Barker and Griffiths, 1972; British Antarctic Survey, 1985) and a little work in the immediate vicinity of South Georgia (Simpson and Griffiths, 1982).

Headland (1982) included a complete bibliography of all the published geological work on South Georgia until 1979. He has also written a complete account of the history of South Georgia (Headland, 1984). The references in this memoir do not constitute an exhaustive bibliography, but all major geological works are cited, and the reference list includes all geological papers on South Georgia published between 1979 and December 1986.

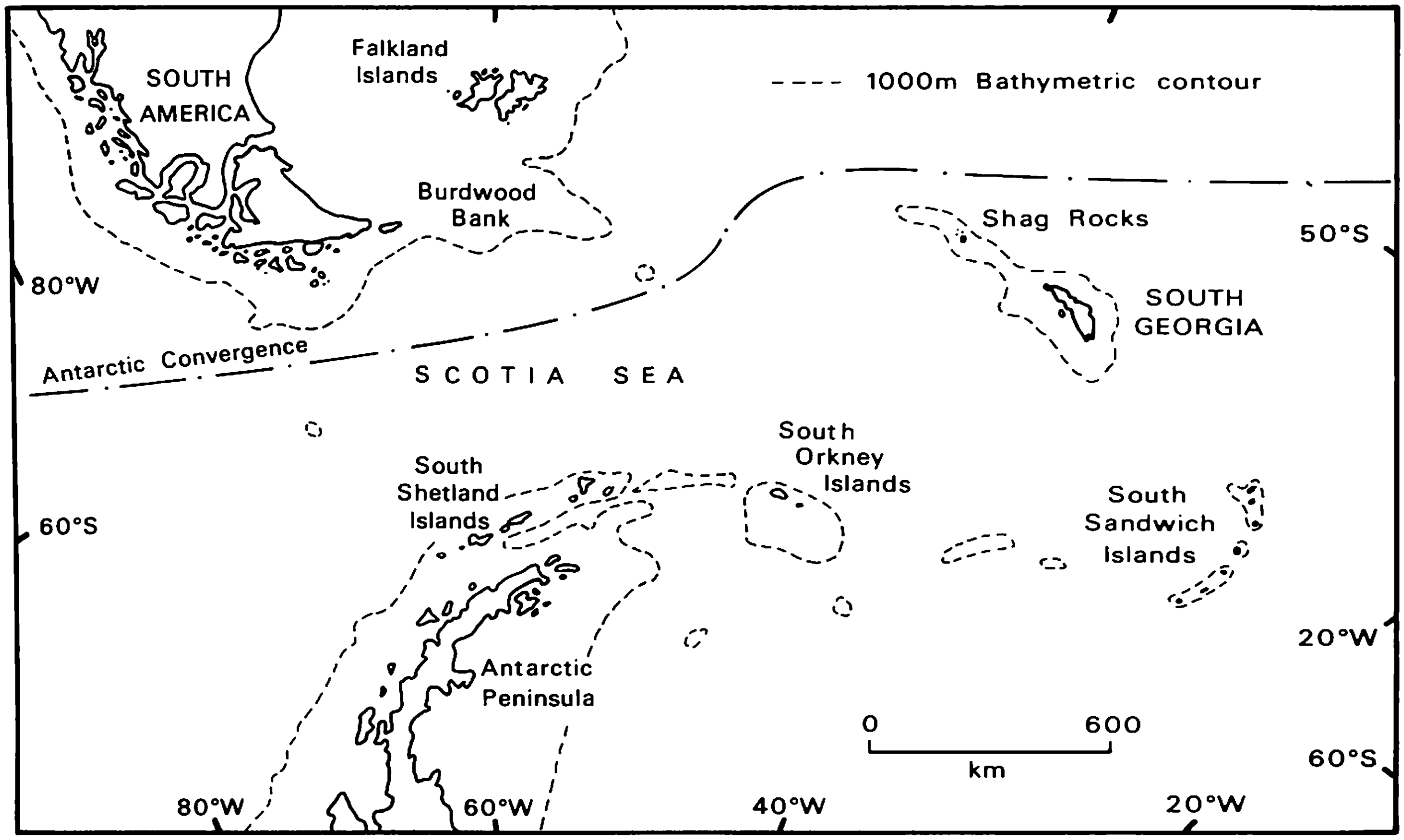


Fig. 1. Location map, showing the position of South Georgia relative to South America and Antarctica.

2 Tectonic setting

Present-day plate configuration

South Georgia sits almost in the centre of a small block of continental crust (*c.* 350 × 200 km) which is roughly rectangular in shape, its long axis oriented north-west–south-east parallel to the axis of the island (Simpson and Griffiths, 1982). The edge of the block is marked by the 500 m isobath and the submarine topography falls away steeply, especially to the north (into the Malvinas Outer Basin) and south (to the abyssal plain of the Scotia Sea). To the west-north-west lies a discontinuous line of small topographically high blocks which form the northern Scotia Ridge, including the Shag Rocks microcontinent (Tanner, 1982*a*), and which connects through Burdwood Bank to the South American continent east of Tierra del Fuego. To the east-south-east, a line of seamounts joins the South Georgia block to the active volcanic arc of the South Sandwich Islands.

The present-day plate tectonic setting of the island is shown in Fig. 2. South Georgia sits in the north-eastern part of the Scotia Plate, which is bounded to the north and south by inactive sinistral transform faults against the South American and Antarctic plates respectively. To the east it is separated by an active spreading ridge from the Sandwich Plate. The western margin of the Scotia Plate is more complex, with the Shackleton Fracture Zone (SFZ) forming the margin against the Shetland and Drake plates in its southern part; the SFZ passes northwards into an inactive or very slow subduction zone, with the Antarctic Plate being subducted eastward below the Scotia and South American plates. The Scotia Plate is complicated, with north–south magnetic lineaments in the eastern and western portions, but east–west striping in the central Scotia Sea, adjacent to South Georgia. Despite the complex anomaly pattern, the whole of the Scotia Plate is clearly of Oligocene or younger age (British Antarctic Survey, 1985) with the oldest anomaly being 10 (*c.* 25 Ma, mid-Oligocene).

The area is almost aseismic; with the exception of the South Sandwich Islands and the few active volcanoes in the South Shetland Islands, there are few tremors. The northern Scotia Ridge in general, and the South Georgia block in particular, are extremely quiet seismically. No earthquakes have been detected on the South Georgia block since 1960 (British Antarctic Survey, 1985).

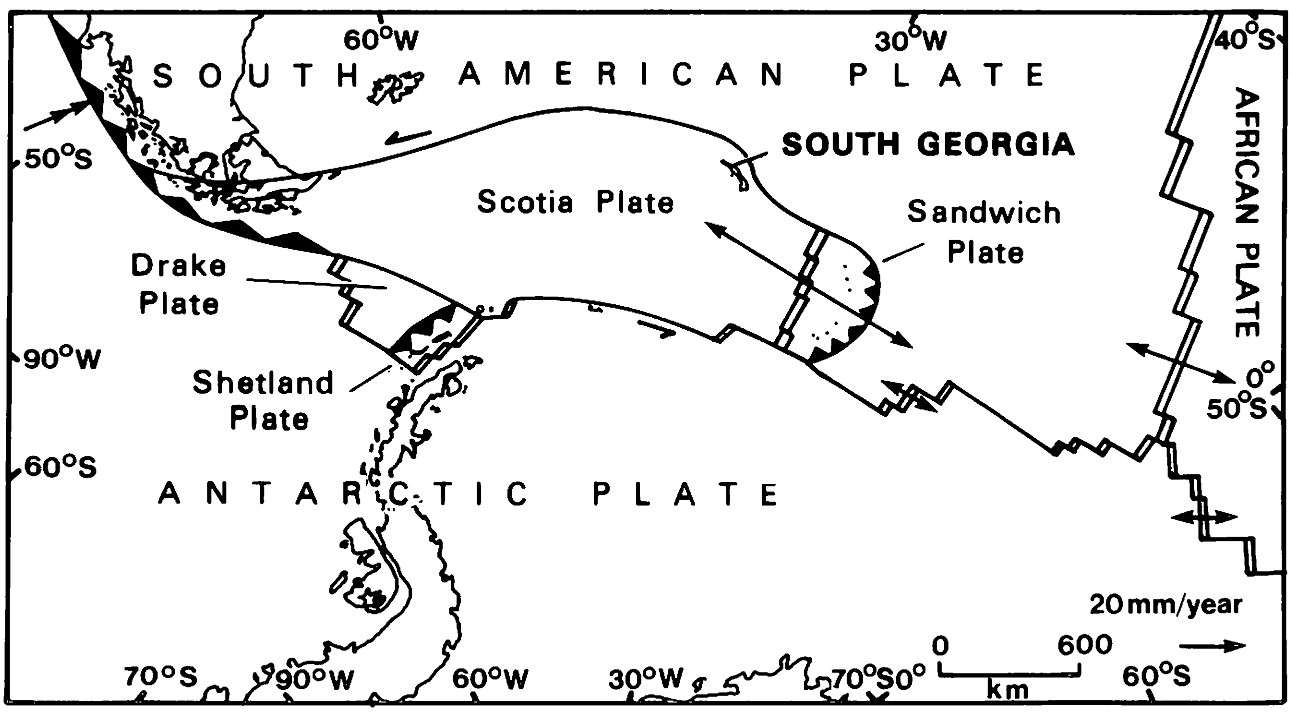


Fig. 2. Present-day plate-tectonic setting of South Georgia (after British Antarctic Survey, 1985).

Mesozoic plate-tectonic evolution

The understanding of the geology of South Georgia is inextricably linked to knowledge of the southern Andes and the Scotia arc. Research has pursued two major avenues: the relative disposition of the various parts of the Scotia arc, and the study of the evolution of the feature as a whole. Although the two topics are interlinked they can be considered separately here, since South Georgia has been used in entirely different ways in each.

Palaeogeographic reconstructions: As the largest and most remote island of the Scotia arc, South Georgia has always been particularly important. Early theories, which regarded the arc as a continuous, locally emergent mountain chain (Nordenskjöld, 1905), were first challenged by Hawkes (1962), who suggested that the islands were disrupted fragments of an essentially rectilinear cordillera joining the two continents. Hawkes' theory was elaborated over the succeeding 20 years and two major areas of disagreement emerged — the original shape of the arc and the position of South Georgia within it. Three major shapes were proposed: rectilinear (Fig. 3a: Hawkes, 1962; Dalziel and Elliot, 1971; de Wit, 1977; Dott and others, 1982), gently cuspate (Fig. 3c: Suárez and Pettigrew, 1976; Dalziel, 1983) and strongly cuspate (Fig. 3b: Barker, 1970; Barker and Griffiths, 1972; Hill and Barker, 1980; King and Barker, in press). Each of these three proposals have two common features, in that South Georgia is placed south of the Burdwood Bank, restored without rotation (Dalziel and others, 1975; Tanner 1982*b*), and the Antarctic Peninsula and South America are placed end-to-end, emphasizing the continuity of terranes through the mountain chain. In contrast, Harrison and others (1979) suggested a major overlap of the Antarctic Peninsula west of South America. Although this reconstruction preserves the position of South Georgia relative to South America, no continuity through the Scotia arc is implied.

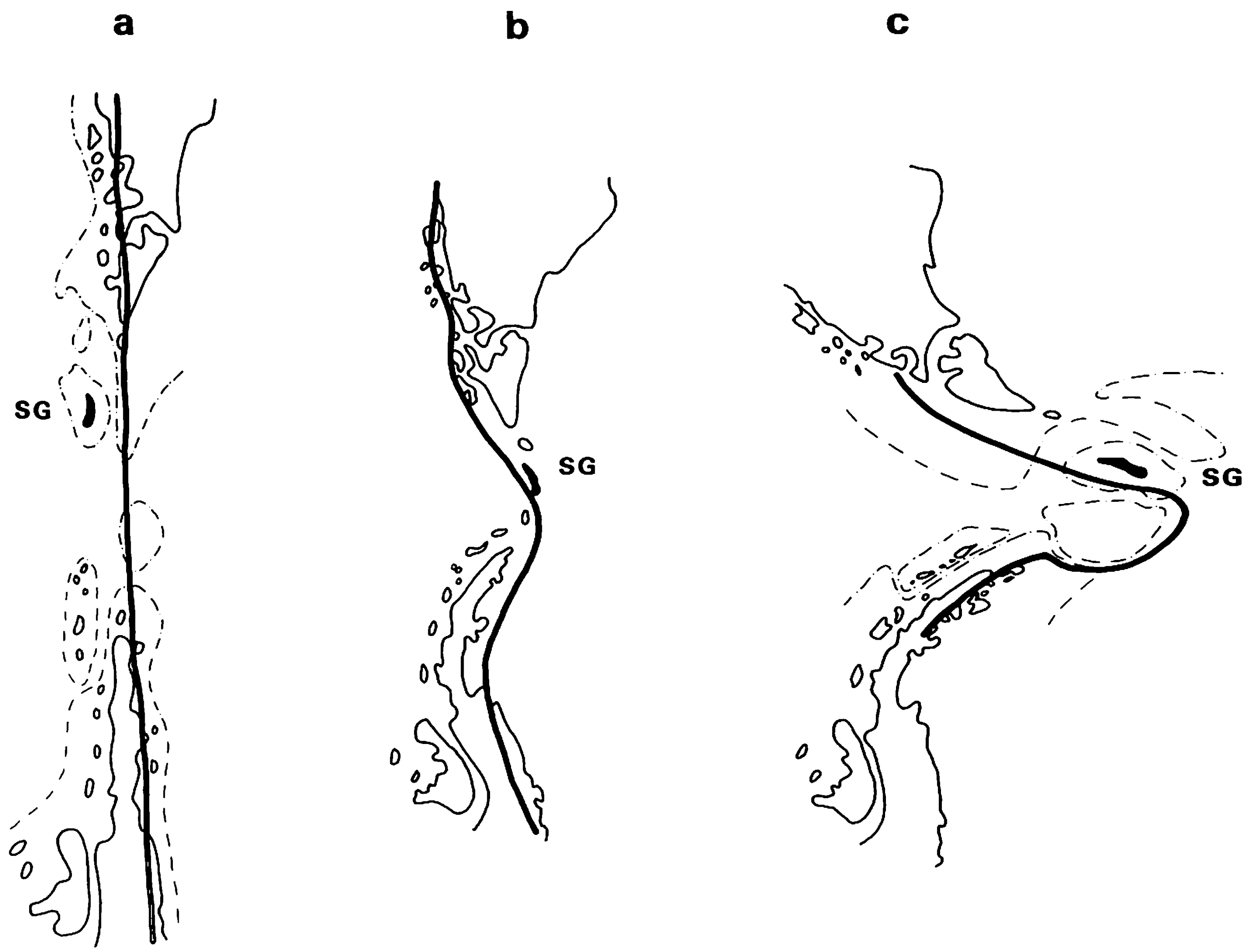


Fig. 3. Alternative reconstructions of the former link between South America and the Antarctic Peninsula showing the position of South Georgia (SG). The geological boundaries chosen by the various authors are not directly comparable, but in each case one boundary is shown to emphasize the continuity of terranes between South America and Antarctica.
a. Reconstruction after Dalziel and Elliot (1971); geological boundary is margin between cordillera and extra-cordilleran basins.
b. Reconstruction after Suárez and Pettigrew (1976); geological boundary is eastern margin of arc terrane.
c. Reconstruction after King and Barker (in press); geological boundary is eastern margin of main arc plutonic suite.

There was also disagreement over the relative position of South Georgia within the Cordilleran belt. Barker and Griffiths (1972) placed the South Georgia block on the Atlantic side of the Cordillera whereas Dalziel and Elliot (1971, 1973) suggested that it represented part of the Pacific margin. This latter suggestion has been largely confirmed by more recent work (Simpson and Griffiths, 1982).

Despite the uncertainties in the reconstruction of the Scotia arc, there is little doubt that disruption was due to the formation of the Scotia Sea and opening of Drake Passage since the mid-Oligocene (Barker and Burrell, 1977). Any further constraints on the pre-Oligocene tectonic evolution of the Scotia arc must come from detailed study of the geology of the component parts of the arc.

Geological evolution: In contrast to the central importance of South Georgia in reconstruction studies, evidence from the island was largely peripheral to the early development of ideas on the evolution of the south-west Gondwana margin. Katz (1973) described the evolution of Tierra del Fuego as a eugeosyncline and used the western Pacific marginal basins as an analogue, but he pursued the idea no further. It was left to Dalziel and others (1974) to put this idea into plate tectonic terms. They recognized all the elements of an island-arc–marginal-basin system which was active along the south-west margin of Gondwanaland from the middle Jurassic to the mid-Cretaceous (Fig. 4). This interpretation results in a division of the Patagonian Andes into three main terranes: island-arc, marginal basin and stable continent.

They interpreted the evolution of the area in four stages:

(a) Creation of a pre-Middle Jurassic metamorphic complex on the margin of Gondwana, perhaps within an accretionary prism. This was dismembered by Middle Jurassic rifting of the continental margin, accompanied by the creation of a regional unconformity and widespread silicic volcanism (Tobífera Formation).

(b) Continued rifting led to complete rupture of the continental crust and the formation of a marginal basin. The floor of the marginal basin is represented by the extensive mafic volcanic rocks of the Rocas Verdes, including the Sarmiento and Tortuga ophiolite complexes. Elongate strips of mafic volcanics are separated by slivers of deformed metasediments.

(c) Uppermost Jurassic–mid-Cretaceous arc volcanism and plutonism, coeval with basin infilling; sedimentation of volcaniclastic material (Yahgan Formation) on the arc side of the basin and silicic detritus on the continental side. On the Pacific side of the marginal basin terrane lies the huge, calc-alkaline Patagonian batholith, intruded between the late Jurassic and the end of the Cretaceous. Dalziel and others (1974) interpreted this as being comagmatic with andesitic volcanism on the south-west margin of the basin (Hardy Formation: Suárez and Pettigrew, 1976).

(d) Closure of the basin in the mid-Cretaceous with deformation of the basin fill, low-grade metamorphism and uplift of the Cordillera.

In Dalziel and others' (1974) seminal paper, little mention is made of South Georgia. However, their model was considerably strengthened when they visited the island and made detailed comparisons (Dalziel and others, 1975). Although South Georgia only preserves part of the total width of the cordillera, the geology can be very closely matched with South America. Over the next five years the model originally proposed by Dalziel and others (1974) was refined, in large part due to detailed work on South Georgia, where access and exposure are better than on the thickly wooded mountains of Tierra del Fuego. Most importantly, pre-existing continental basement and a major ophiolite were recognized on South Georgia (Bell and others, 1977; Storey and others, 1977). These discoveries, coupled with the basin-fill (Dalziel and others,

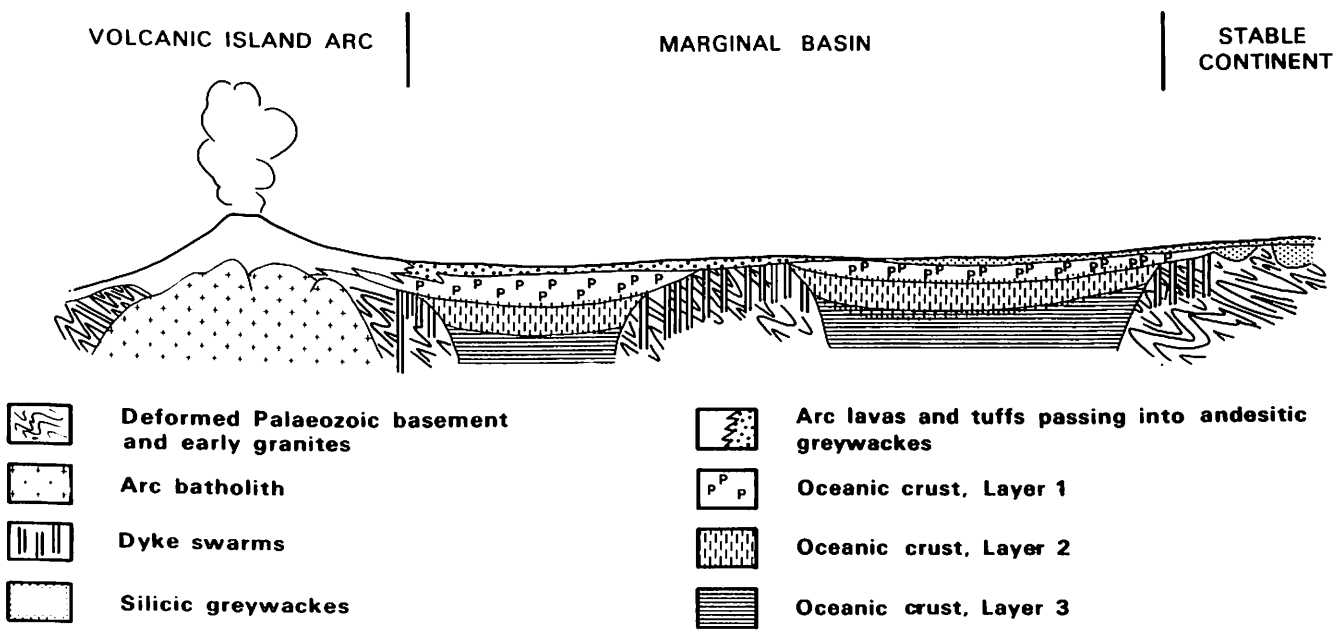


Fig. 4. Restored diagrammatic section across the southern Andes in early Cretaceous times, showing the three main terranes (after Dalziel and others, 1974).

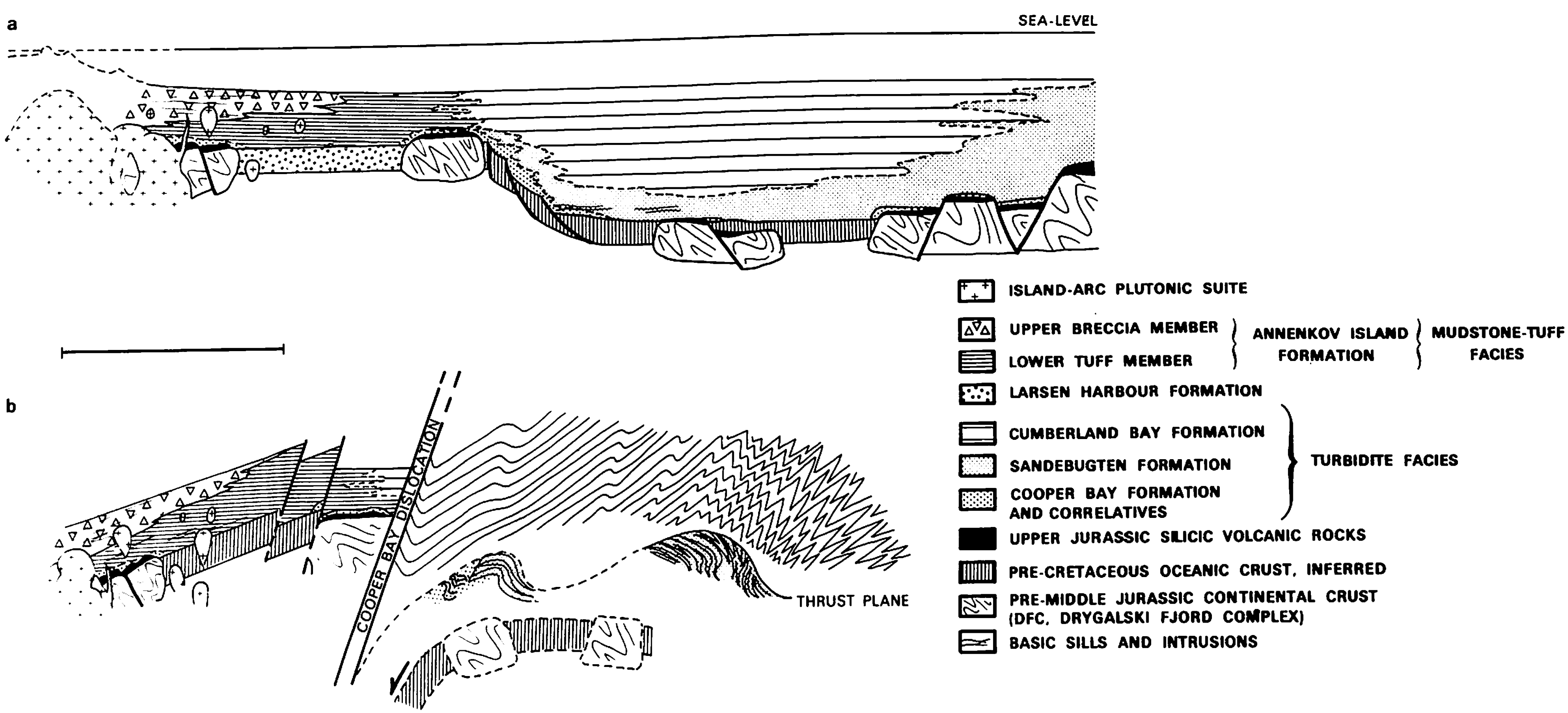


Fig. 5. Schematic cross-section of the South Georgia area:
a. at the close of sedimentation late in the early Cretaceous, and b. after the Andean (mid-Cretaceous) deformation. Both profiles are drawn along an approximate line through Annenkov Island and Barff Point; they are not to scale but the bar represents c. 25 km. Lateral relationships between formations have been considerably telescoped for clarity and the island-arc plutonic suite, for example, is situated at a considerably greater distance from the Larsen Harbour Complex than is shown here (after Tanner and others, 1981, fig. 25).

1975) and island-arc terranes (Suárez and Pettigrew, 1976) already recognized meant that both arc and basin were represented on the island, with only the former continental margin missing. Tanner and others (1981) produced the most detailed model for the geological evolution of South Georgia (Fig. 5). This model is specifically for the South Georgia portion of the back-arc basin, but also draws on evidence from South America, especially with regard to the continental margin.

Table II shows the relationship of the major rock units; each of these units is described and discussed in turn in the following sections.

Table II. Summary of the lithostratigraphy of South Georgia

SOUTH-WEST									*NORTH-EAST*
ANNENKOV ISLAND FORMATION (?uppermost Jurassic–Aptian/Albian)		Upper Breccia Member Lower Tuff Member[1]	COOPER BAY DISLOCATION ZONE	DUCLOZ HEAD FORMATION (Age unknown) Coastal Member[2]	Fault	Inland Member[1]	Fault	CUMBERLAND BAY FORMATION (?Neocomian–Aptian/Albian)	Facies Change	Barff Point Member
		? ? ? ? ?						Thrust fault		
LARSEN HARBOUR COMPLEX (Upper Jurassic)	? Fault	DRYGALSKI FJORD COMPLEX (?Triassic–Upper Jurassic)		COOPER BAY FORMATION[2] (Age unknown)				SANDEBUGTEN FORMATION[2] (?Mesozoic)		

Notes

1. Lower Tuff Member of the Annenkov Island Formation and Inland Member of the Ducloz Head Formation are lithologically equivalent.
2. Sandebugten Formation, Cooper Bay Formation and Coastal Member of the Ducloz Head Formation are lithologically equivalent.

3 Drygalski Fjord Complex

Introduction

The Drygalski Fjord Complex (DFC) is an area of igneous, metamorphic and metasedimentary rocks comprising deformed paragneisses (Salomon Glacier Formation), metamorphosed sedimentary rocks (Cooper Island and Novosilski Glacier formations), large gabbro and small granitic plutons, and migmatites. It represents a fragment of pre-Jurassic continental crust (basement) that was intruded by a wide variety of mafic and felsic plutons prior to and during the early stages of formation of the island-arc–back-arc-basin system of South Georgia.

The complex forms part of the Salvesen Range, the central range of southern South Georgia, with jagged peaks rising to over 2000 m and culminating in Mount Carse (2331 m).

Although both igneous and metamorphic rocks were known from this area since Filchner's German South Polar Expedition (Heim, 1912), it was Trendall (1959) who first mapped the area and referred to it as the "South-Eastern Igneous Complex". More extensive mapping was carried out in 1974–76 (see Bell and others, 1977; Storey and others, 1977; Tanner and Rex, 1979; Storey and Mair, 1982; Storey, 1983*a*) and the unit was renamed the Drygalski Fjord Complex (Storey, 1983*a*). The following is a summary of the DFC, the details of which can be obtained from Storey (1983*a*).

Stratigraphic and age relations

The DFC is separated from the deformed basin-fill sediments of the Cumberland Bay and Cooper Bay formations by a major shear zone, the Cooper Bay dislocation (Tanner, 1982*b*; formerly the Mylonite Zone of Storey and others, 1977). This zone is well-exposed at the south-east corner of the island, trending north-east in a series of high ridges inland from Cooper Bay. Farther north it is not exposed, but the dislocation zone is inferred to follow Spenceley Glacier. On its south-western margin, the DFC is separated from the ophiolitic rocks of the Larsen Harbour Complex by a prominent topographic low occupied by Drygalski Fjord, and Risting and Novosilski glaciers. This is interpreted as a fault zone. The complex probably extends south at least as far as Clerke Rocks (Tanner, 1982*b*).

The age of the various lithologies is not well constrained. The sedimentary and metasedimentary rocks are intruded by granites and gabbros which have yielded early Jurassic Rb–Sr and K–Ar ages (Tanner and Rex, 1979). These are the oldest known rocks on South Georgia, and may constitute the basement terrane. K–Ar mineral ages for the migmatite, granite and dykes fall within the 120–149 Ma range, which indicates that magmatism and high geothermal gradients may have existed during much of the Jurassic.

Sedimentary and metasedimentary rocks

Sedimentary and metasedimentary rocks predate all the intrusive rocks, and form about 30% of the total outcrop. They vary from sediments with identifiable angular clastic grains to intensely deformed paragneisses. There are three spatially distinct formations which do not have exposed boundaries but are believed to represent different structural and metamorphic levels of a single sequence. The deepest levels crop out in the southern part of the main island, whereas in the northern part and on Cooper Island high-level magmas were intruded into clastic and volcaniclastic rocks.

Salomon Glacier Formation (SG): The metasedimentary rocks of the Salomon Glacier Formation (defined by Storey, 1983*a*) range from fine-grained quartz-andesine-perthite-biotite-paragneiss showing relict sedimentary structures (cross lamination, disturbed bedding, slump folds and load and flame structures), to intensely deformed banded gneisses and layered migmatites (Fig. 6) regionally metamorphosed up to the amphibolite facies.

Deformation of the SG varies locally. In the Trendall Crag area weakly deformed paragneisses are folded by isolated close to open sub-vertical folds and cut by ptygmatic quartz-feldspar veins. Around the Salomon and Bogen glaciers the gneisses are more intensely deformed, with evidence of three phases of folding associated with syn-tectonic metamorphism to amphibolite facies and separation of migmatitic segregations (see Story, 1983*a*).

Cooper Island Formation (CI): A flat-lying, inverted sequence, *c.* 50 m thick, of laminated mudstone, siltstone and massive sandstone crops out on the north-eastern promontory of Cooper Island. This, together with lenses of banded, hornfelsed metasediment at scattered localities throughout the island, forms the Cooper Island Formation (defined and described by Storey, 1983*a*). Convolute lamination, cross lamination and graded beds are well preserved in laminated units. Bottom structures are also well developed (Fig. 7) and include load and flame structures, rill and groove casts and prod marks. Rare plant-stem impressions, vermicular trace fossils and irregular burrows are preserved in

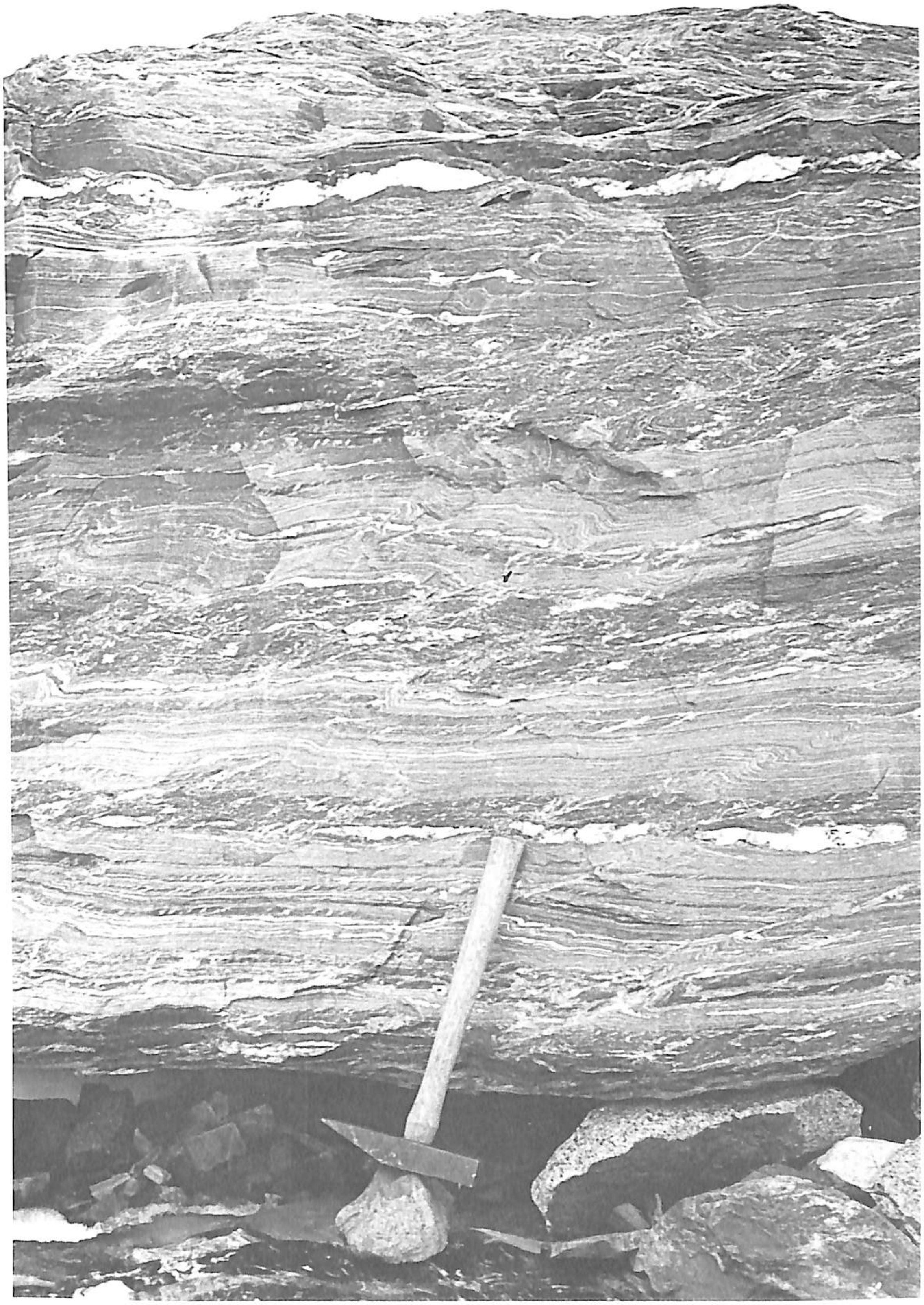


Fig. 6. Folded layered migmatites of the Salomon Glacier Formation. The hammer shaft is 35 cm long.

the mudstone horizons. The sandstones are moderately to poorly sorted immature arkosic greywacke and arkosic arenite (*sensu* Pettijohn and others, 1972) derived from a granitic, metamorphic and volcanic terrane.

The sedimentary rocks exposed on the north-eastern promontory probably lie on the inverted limb of a major fold of unknown orientation. The slight variation in the dip of the flat-lying beds is the result of open folding;

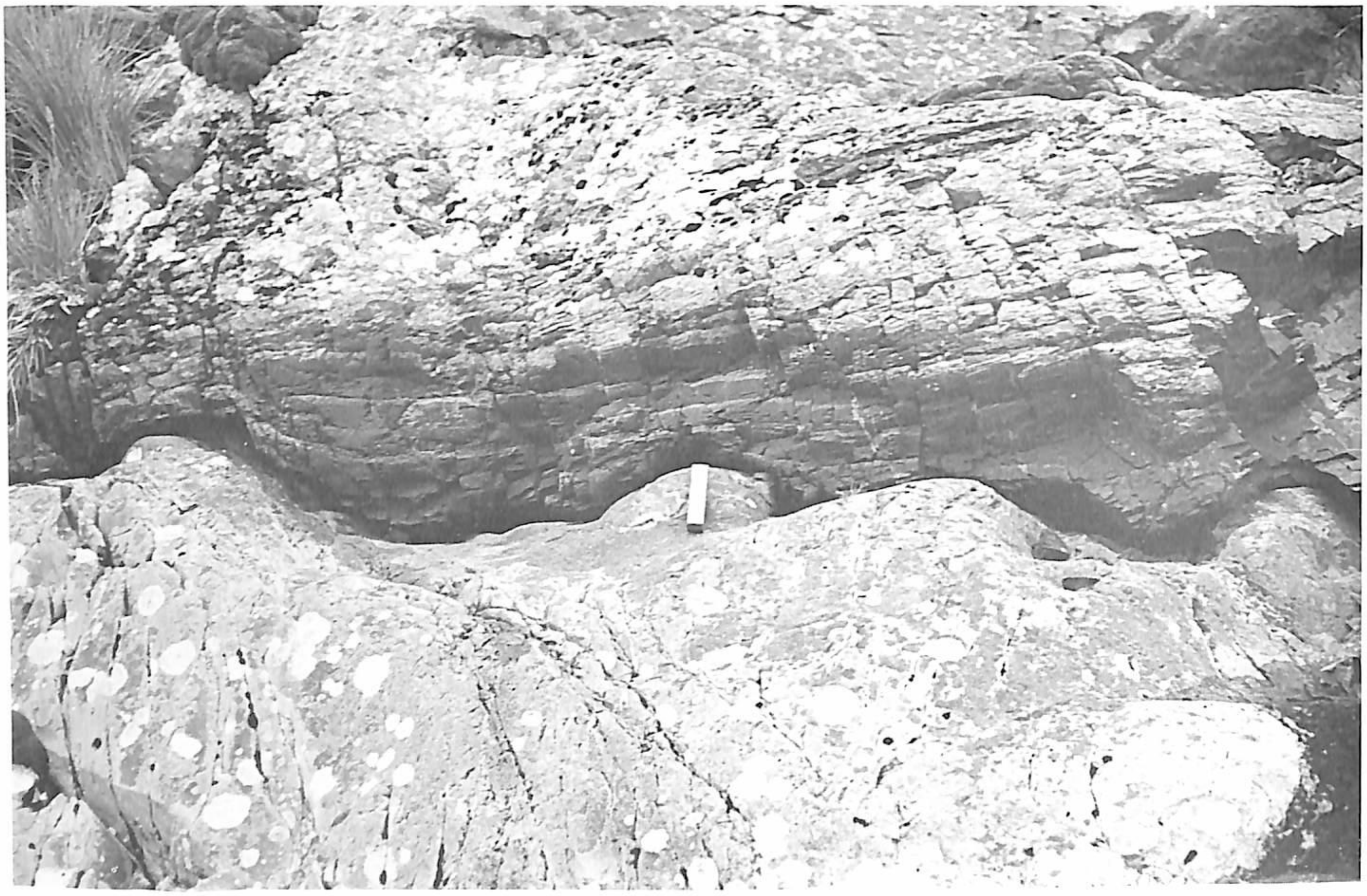


Fig. 7. Overturned load structures within the Cooper Island Formation. The scale is 10 cm long.

late small-scale faulting and brittle deformation considerably disrupt the bedding. Secondary mineral assemblages indicate a low grade of regional metamorphism although the banded hornfelses reflect a contact thermal effect up to the biotite-cordierite zone of the hornblende-hornfels facies, with retrogression to a chlorite-epidote assemblage.

Novosilski Glacier Formation (NG) (defined and described by Storey, 1983*a*): Epiclastic and volcaniclastic sandstones, with interlaminated mudstone and shale, and porphyritic felsites of this formation crop out in a number of scattered localities in the Novosilski, Spenceley and Risting glacier basins in the northern part of the complex, and within the northern part of the Cooper Bay dislocation zone. Although the true relationship between the felsites and epiclastic sedimentary rocks is not exposed, mainly due to the strong imprint of the mylonitic fabric in the dislocation zone, petrographic studies suggest that they are gradational and part of the same sequence.

The sedimentary rocks of the NG within the complex were metamorphosed under low-grade conditions. Within the dislocation zone they were metamorphosed up to biotite-grade greenschist facies and were deformed prior to intrusion of the igneous rocks, mylonitization and subsequent deformation.

Discussion: The clastic sedimentary rocks of the Novosilski Glacier and Cooper Island formations are petrographically similar. They were both derived from a granitic and granite-gneiss terrane and from felsic volcanic rocks. As the hornfelsed metasediments of the CI are petrographically similar to the paragneisses of the Salomon Glacier Formation it is also possible that the latter are the metamorphosed equivalents of the clastic sedimentary rocks.

Thus the sedimentary and metasedimentary rocks are probably part of the same succession, subjected to varied degrees of deformation and metamorphism prior to emplacement of the igneous rocks; they now form part of a pre-Jurassic basement to South Georgia. Although there is little evidence to constrain the tectonic setting of these rocks, they are similar to the late Palaeozoic–early Mesozoic quartzo-feldspathic sediments of the Trinity Peninsula Group (TPG) in the Antarctic Peninsula. The TPG rocks probably represent part of a fore-arc region (Storey and Garrett, 1985), and it is possible that the rocks from South Georgia were formed also in a pre-Jurassic fore-arc.

Igneous rocks

Magmatic activity commenced in the early Jurassic (Tanner and Rex, 1979) with emplacement of a subalkaline tholeiitic magma. Differentiation of this magma produced a wide variety of layered gabbros, and intermediate and granitic rocks. Continued emplacement of mafic magma formed a mafic dyke suite (with chilled

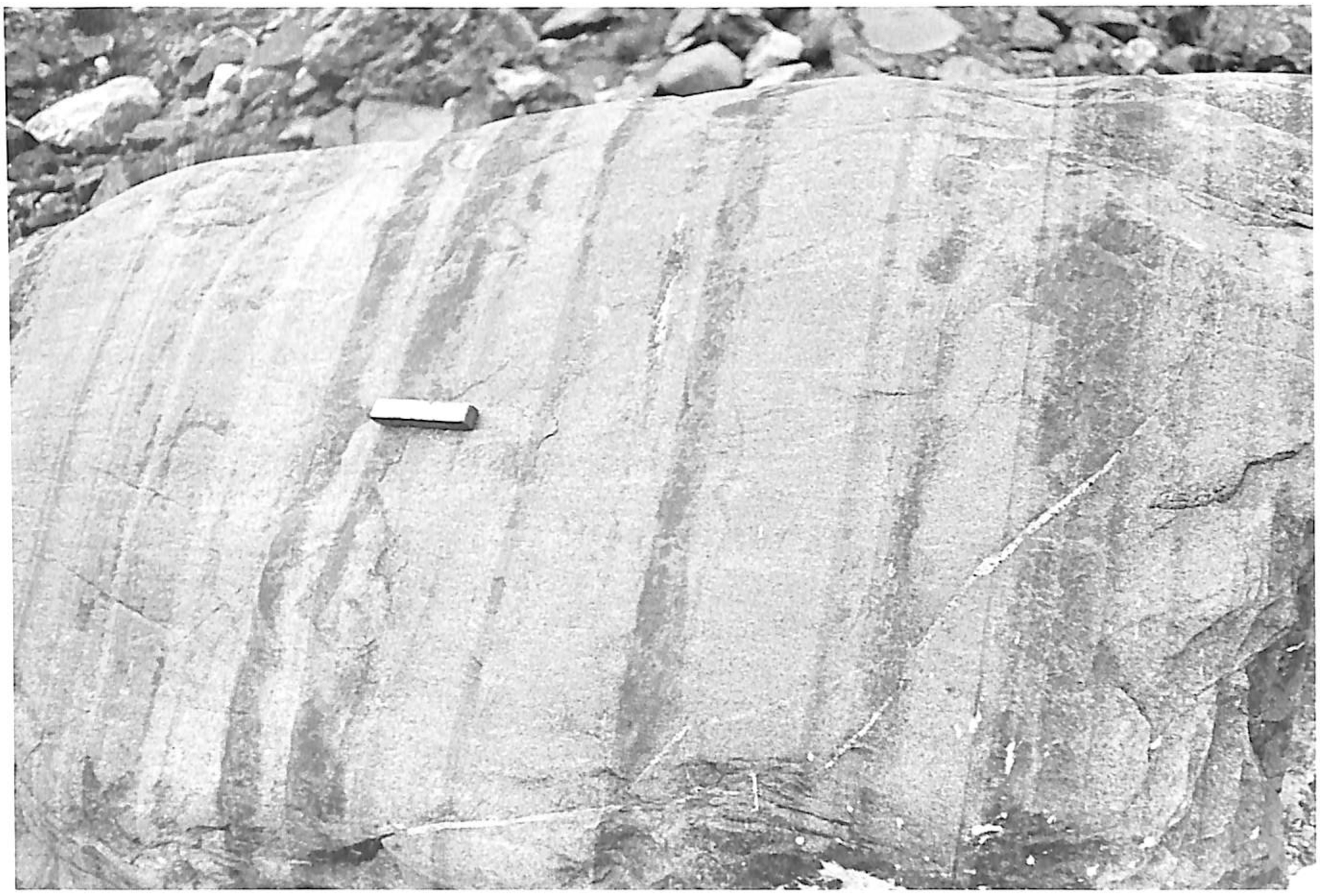


Fig. 8. Cumulate layering within the gabbros, Bogen Glacier. The scale is 10 cm long.

margins) cutting the gabbro and granitoid rocks. A migmatitic aureole surrounds the plutonic rocks as a result of the emplacement of large volumes of mafic magma into the metasedimentary rocks. The mafic, granitoid and metasedimentary rocks are broken up, injected and net-veined by anatectic granitoid veins and dykes, and by the emplacement of intrusive breccias.

Gabbros: Gabbroic rocks vary in composition from feldspathic leucogabbro to ultramafic rocks and are the most abundant igneous rock of the complex. They were emplaced from the early Jurassic onwards and range from small intrusive bodies to large irregular plutons. The larger plutons commonly exhibit cumulate layering (Fig. 8) and are mainly exposed in the south of the complex. Most are medium-grained rocks of variable composition (troctolite, olivine-gabbro, hornblende-pyroxene-gabbro, two-pyroxene-gabbro, norite, leucogabbro and metagabbro; Storey, 1983*a*). Many of the larger plutons, as well as displaying rhythmic layering, are composite bodies: intrusive contacts are present and enclaves of variable grain size and composition are formed within some intrusions. Ultramafic inclusions (lherzolite, peridotite and picrite) are present within massive recrystallized leucogabbro in the Hamilton Bay area. Coarse gabbroic pegmatites cut layered and massive gabbro, and coarse-grained crystal segregations of feldspar and ferromagnesian minerals occur occasionally.

Layering within the gabbro is rhythmic and commonly steeply inclined and, to the west of Bogen Glacier, shows large-scale folding. Layer thickness varies from less than 1 cm to massive bands up to 1 m thick. Sharp and gradational contacts separate successive layers, which are strikingly regular. North-east of Hamilton Bay there is an area of irregular, commonly discontinuous layering, folding, unconformities and cross bedding within a layered gabbro; large angular blocks of layered gabbro are also found within more massive gabbro.

Diorites and quartz-diorites: Dioritic rocks are present in the Trendall Crag area and on Cooper Island and form a small proportion of the mafic rocks within the DFC. They occur along gabbro-granite contact zones, form inclusions up to 50 m wide within the Trendall Crag granodiorite and form dykes within the metasediments to the east of Bogen Glacier. On Cooper Island granitic veins occur in a characteristic reticulate network (Fig. 9) within the diorite; the subhorizontal granitic veins have sharp lower contacts and irregular fretted upper margins. The veins may have formed by *in situ* differentiation of the mafic magma during crystallization.

The dioritic rocks are medium-grained, dark-coloured, biotite- and hornblende-rich rocks which often contain quartz and plagioclase phenocrysts, and rounded quartz ocelli rimmed by opaque minerals and amphibole laths. The ocelli are usually formed along granite-gabbro contact zones.

Fig. 9. Granitic veins within the Cooper Island diorite. The hammer shaft is 60 cm long.

Granitoids: Granitoids form a small but significant part of the DFC and include two large mappable units: the Trendall Crag granodiorite and the Cooper Island granophyre (Storey, 1983*a*). Throughout the remainder of the complex trondhjemite, tonalite, granodiorite and alkali-granite dykes intrude and form lenses within the gabbros, mafic dykes and metasedimentary rocks. In the northern part of the complex, porphyritic (commonly flow-banded) rhyolite and dacite, and porphyritic microgranites form dykes and lenses within the numerous mafic dykes and sediments. Contact relationships between the granitic dykes and lenses and the surrounding rocks are often complex; in most cases the dykes intrude and have chilled margins, but gradational contacts do exist and back veining may occur.

The Trendall Crag granodiorite is a composite body of variable composition (tonalite, granodiorite and granite). It intrudes and assimilates the gabbro plutons with the formation of marginal ocellar diorites, net-veins, agmatites and dioritic assemblages.

The Cooper Island granophyre is a pale grey leucocratic trondhjemite and tonalite with a characteristic granophyric texture and up to 5% ferromagnesian minerals, which are commonly replaced by epidote, chlorite, actinolite and muscovite. It intrudes and assimilates the Cooper Island Formation and has a gradational contact with the diorite on Cooper Island. Intrusive breccia, which is not seen in the granophyre, was emplaced within the hornfelsed sediment and gabbro, and may represent a marginal, more volatile-rich phase of the granophyre.

Mafic dykes: Mafic dykes (Fig. 10) were intruded at various stages during the history of the complex. Most dykes were intruded prior to, and were affected by, the migmatization event, although a small proportion were intruded during and after migmatization. The concentration of dykes increases towards the western margin of the complex, where they form up to 80% of the outcrop. The dyke rocks are often porphyritic and vary in composition from dolerite to quartz-diorite. Their degree of alteration varies and the most altered ones may have acted as conduits for migrating hydrothermal fluids.

The dykes generally occur as single units 1–2 m wide but dykes up to 10 m wide were recorded. They are mainly straight-sided and are continuous along strike. Thinning and displacement of the dykes does occur, however, and large dykes may abruptly thin and continue as one or more narrow veinlets, or may die out completely. Multiple dykes are present in the Trendall Crag area with up to five separate intrusions emplaced along the centre of the preceding dyke. Although chilled margins are generally developed, trains of angular and partially assimilated mafic dyke fragments in primary quartz-gabbro and diorite indicate that some of the

Fig. 10. Mafic dykes intruding the Trendall Crag granite. The scale is given by the ice axe and rucksack, lower right.

Fig. 11. Syn-plutonic dyke fragments within a massive gabbro, Bogen Glacier. The scale is 10 cm long.

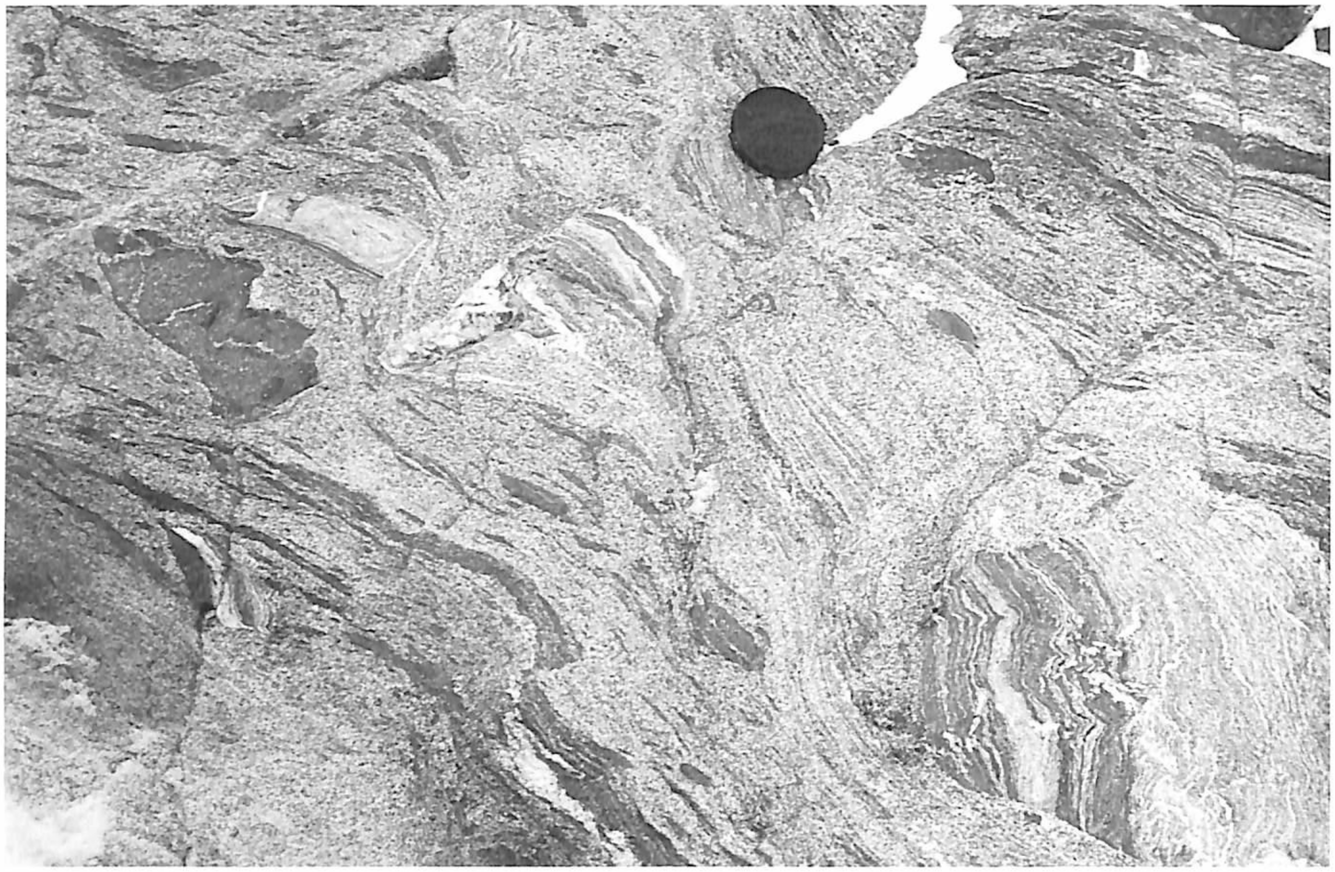


Fig. 12. Fragments of layered migmatite and mafic dyke in a granitic neosome, Bogen Glacier. The lens cap is 5 cm in diameter.

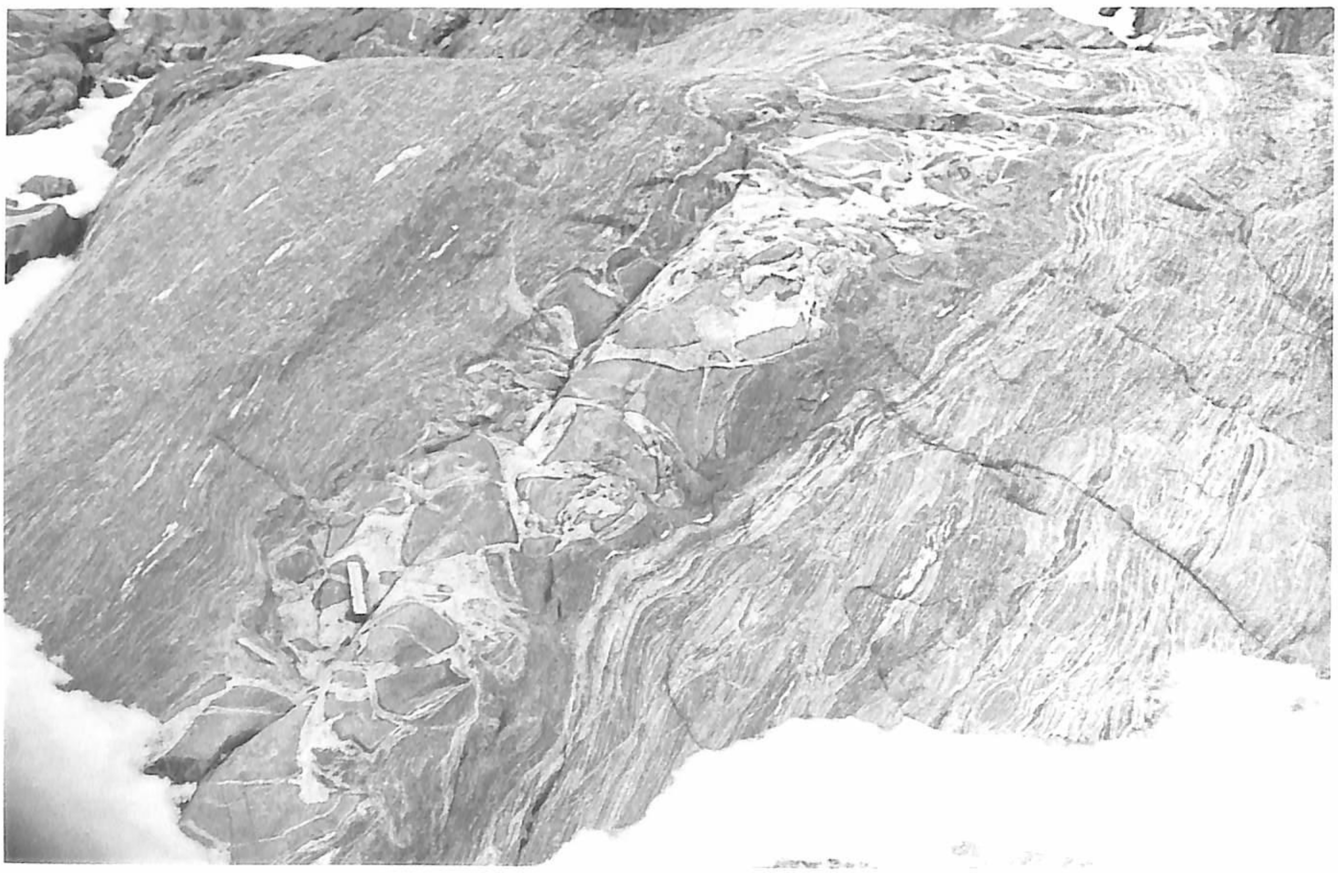


Fig. 13. Mafic dyke disrupted by agmatitic veins, Bogen Glacier. The scale is 10 cm long.

dykes were emplaced contemporaneously with the plutonic rocks (Fig. 11).

A statistical analysis of 681 dyke orientations shows trends which are mostly bimodal with north-west and north-east orientations. The north-easterly dykes have azimuths ranging from 047°–073° while the north-westerly set vary from 283°–329°. In the northern part of the complex a third dominant mode is orientated north–

Fig. 14. Angular mafic fragments cut by granitic veins in an intrusive breccia, Storey Glacier. The hammer shaft is 60 cm long.

Fig. 15. An intrusive breccia containing enclaves of different lithologies, Cooper Island. The hammer shaft is 80 cm long.

south. The cross-cutting relationships of the dykes are complex and although the north-east set is mostly younger than the north-west set, variations do occur.

Migmatites: A heterogeneous migmatite complex cut by aplite, pegmatite and late mafic dykes surrounds the gabbro and granitic plutons and is best exposed east of Bogen Glacier. The mafic, felsic and sedimentary rocks are broken up, injected and net-veined by a fine-grained granitic neosome, with angular and partially assimilated gabbroic and mafic dyke enclaves enclosed within a granitic matrix. All gradations exist from paragneiss, layered migmatite and mafic dykes cut by irregular veins of migmatitic granitic neosome (Fig. 12) to large areas (up to 10 m wide) of medium-grained granite (neosome) with nebulitic gneissose and mafic enclaves, and relict ghost structures. Agmatitic structures occur extensively throughout the Salomon Glacier Formation gneisses, where angular fragments of paragneiss are brecciated by relatively narrow veins of pale-coloured medium-grained granitic and aplitic neosome. The migmatitic granites vary in composition from quartz-diorite and tonalite to granodiorite and granite and show great variation in grain size and texture.

The mafic dykes display a complex range of migmatitic structures; all gradations exist from mafic dykes with cross-cutting veins, to trains of angular fragments disrupted by agmatitic veins (Fig. 13) and to remnant trains of partially assimilated enclaves. The granitic veins commonly preferentially invade the dykes and leave the country rock intact. In some cases there has been complete disruption of the mafic intrusion producing irregular areas of bimodal mafic-felsic breccia, with angular and sub-rounded partially assimilated mafic fragments set in a granitic neosome. The mafic fragments may be entirely derived from a single phase (Fig.14) or as is commonly the case, the breccia contains a wide variety and proportion of different lithologies. The enclaves vary in shape from angular to sub-rounded and include gabbro, amphibolite, mafic dyke, metasediment and occasional granitic fragments. These breccias (Fig. 15) have been termed *intrusive breccias* (Storey, 1983*a*) as it is believed that they were emplaced as breccias with mixed-provenance enclaves. They contrast strongly with the *in situ* migmatization, brecciation and net-veining.

The migmatitic complex was formed by partial melting of the metasedimentary rocks, emplacement of intrusive breccias, interaction of contemporaneous mafic and felsic magmas, and by reactivation and mobilization of earlier granitic phases. The complex represents a period of high geothermal gradients resulting probably from the emplacement of large volumes of mafic magma to a high level in the crust during the development of the marginal basin.

Geochemistry of the igneous rocks

Major- and trace-element analyses of 41 mafic dyke rocks, 48 granitic rocks and 62 mafic plutonic rocks from the DFC were carried out by XRF at the University of Birmingham (Storey, 1983*a*). The mafic dykes and gabbros have a similar chemistry (Fig. 16) and were derived from the same magmatic source, a sub-alkaline tholeiitic magma. The mafic dykes are olivine- and quartz-tholeiites and the gabbros are sub-alkaline olivine-tholeiites. They show a marked iron-enrichment trend on an AFM diagram (Fig. 17) and fall within the tholeiitic field. The dykes, diorites and quartz-diorites are more fractionated than the gabbros.

There is a wide range in chemical composition of the granitic rocks within the complex (Fig. 18). With the exception of TiO_2, Al_2O_3, MnO, Fe_2O_3, MgO and P_2O_5, which decrease with increase in SiO_2, the major and trace elements do not follow well-defined trends. There is considerable variation in K_2O, Rb, Ba, and Sr; many of the samples are dacitic in composition with K_2O <0.25% whilst others, commonly within the same granitic group, have up to 5%. There is a wide variation in Zr, Y and Nb contents; some, in particular the Cooper Island granophyre, the porphyritic felsites and some granitoid dykes, are enriched in Y, Nb and Zr (Fig. 18) and have equivalent Y/Nb and Ce/Y ratios to the mafic rocks. These characteristics, which are similar to within-plate granite (Pearce and others, 1984), suggest that they are derived by differentiation of the mafic rocks. The remaining dykes, lenses, agmatites and migmatitic granites have lower Y, Zr and Nb contents, high Ce/Y ratios and show chemical characteristics transitional between the above mafic differentiates and continental calc-alkaline rocks. They probably formed by partial melting or remobilization of the continental basement during emplacement of the mafic rocks. The rocks with intermediate values (some of the Cooper Island granophyres) may have formed by differentiation of the mafic magma contaminated by assimilation of the meta-

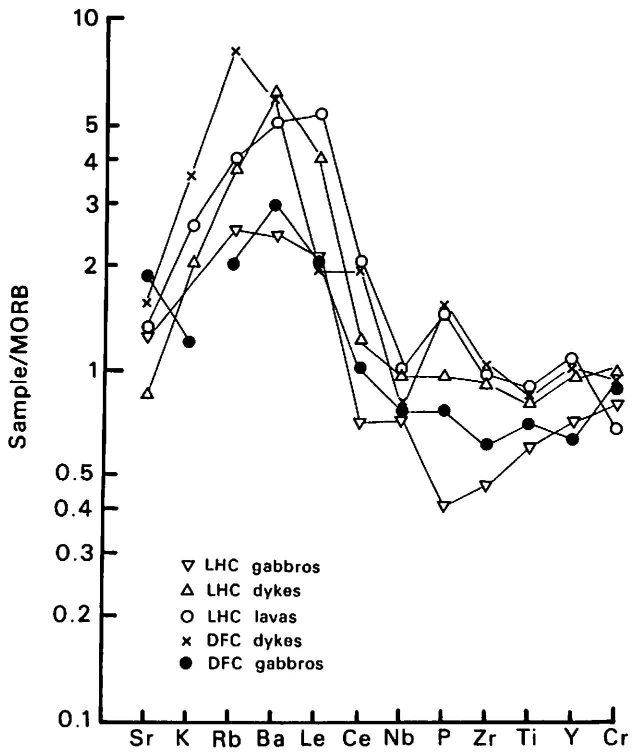
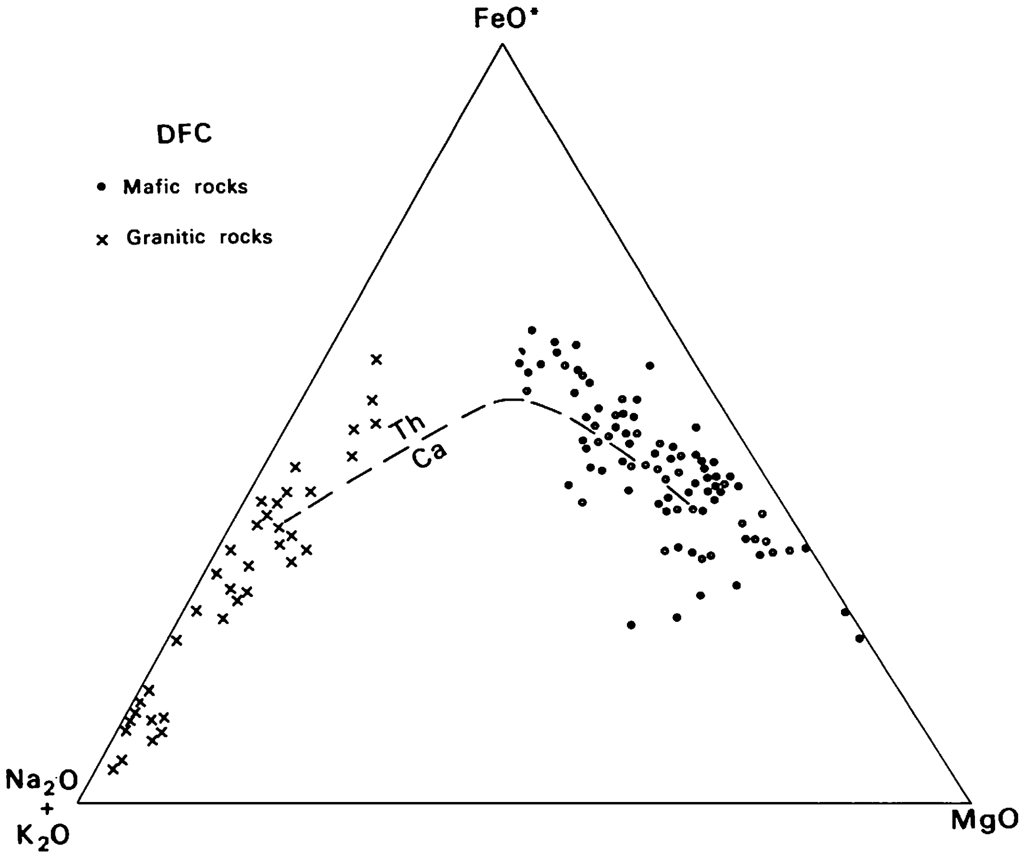


Fig. 16. Average analyses of selected major and trace elements of mafic rocks of the Drygalski Fjord Complex (DFC) and Larsen Harbour Complex (LHC). The analyses are normalized to mid-ocean-ridge-basalt (MORB) values of Pearce (1980).



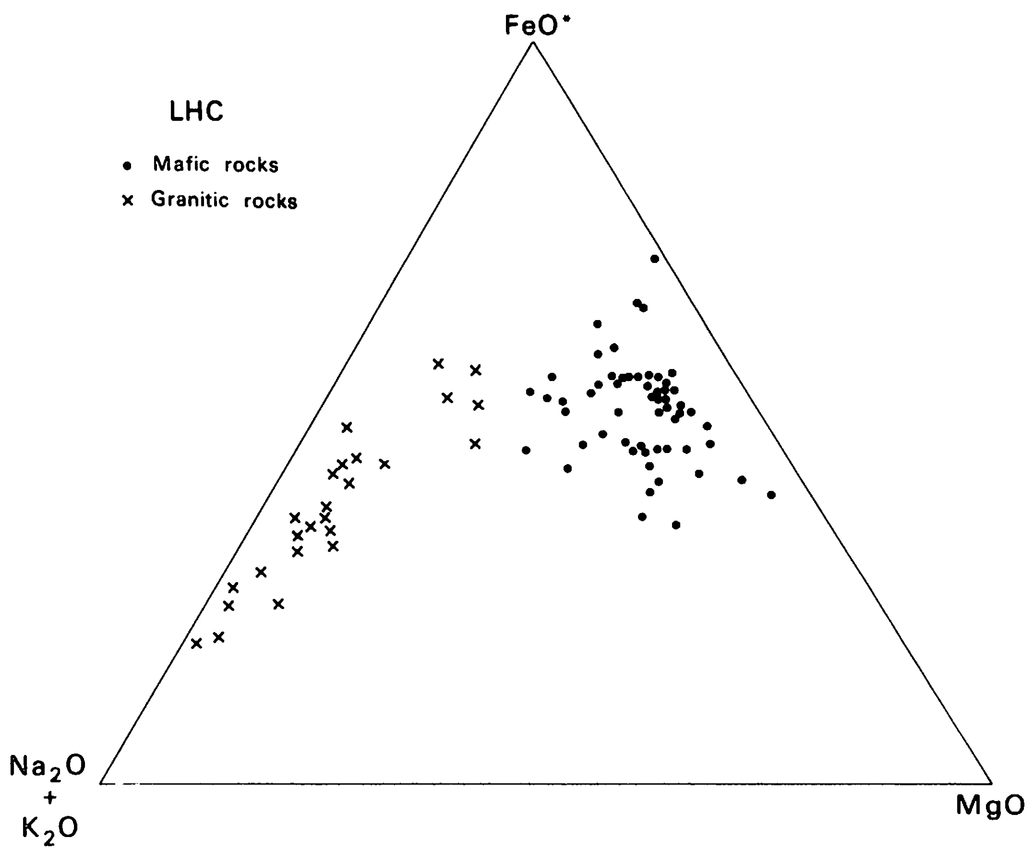


Fig. 17. AFM diagrams of the mafic and granitic rocks of the Drygalski Fjord Complex (DFC) and Larsen Harbour Complex (LHC). The dashed line on Fig. 17a separates the tholeiitic (Th) and calc-alkaline (Ca) fields.

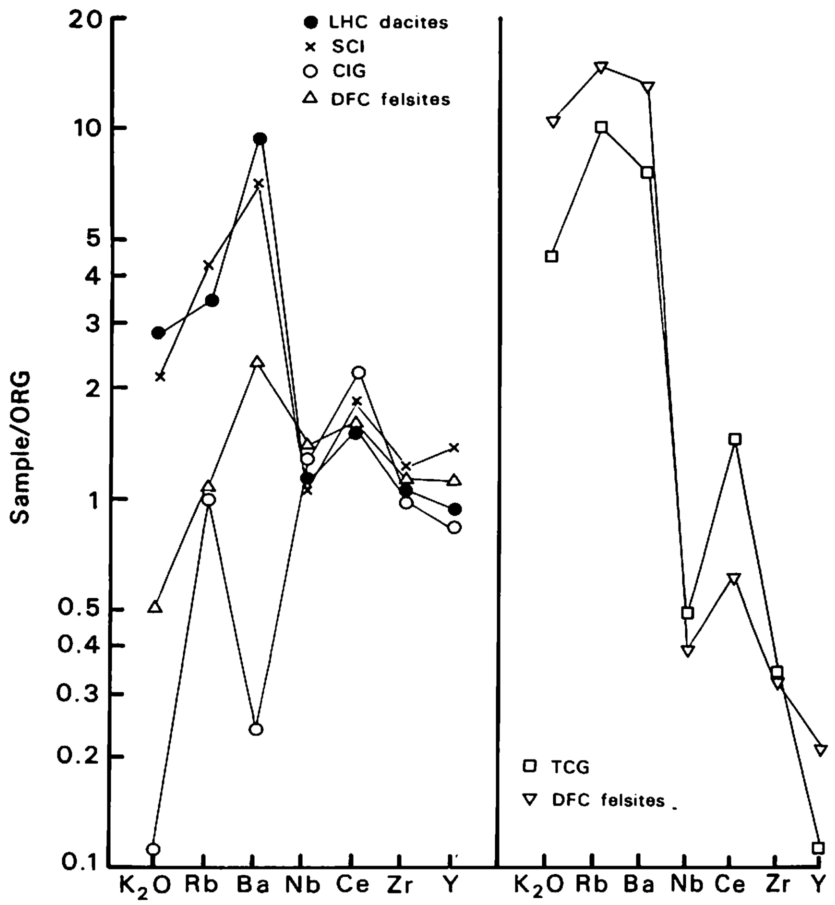


Fig. 18. Average analyses of selected major and trace elements of the granitic rocks of Drygalski Fjord Complex (DFC) and Larsen Harbour Complex (LHC). SCI, Smaaland Cove intrusion; CIG, Cooper Island granophyre; TCG, Trendall Crag granodiorite. The analyses are normalized to ocean-ridge-granite (ORG) values of Pearce and others (1984).

sedimentary rocks. The Trendall Crag granodiorite has characteristics most typical of calc-alkaline rocks and is similar to some plutonic rocks of the Antarctic Peninsula. The initial $^{87}Sr/^{86}Sr$ ratio is 0.7086 which, although high for mantle-derived rocks, precludes a single origin by partial melting of the metasedimentary rocks. These may be related to the complex magma generation processes associated with calc-alkaline rocks formed by subduction.

It is thus concluded that the mafic rocks are continental tholeiites, that all are derived from the same magmatic source and are enriched in LIL elements compared to MORB. The granitic rocks are variable in origin. Some are cogenetic with the mafic magma and formed by differentiation of the tholeiitic magma, whereas others formed by partial melting of the metasedimentary rocks during emplacement of the mafic rocks; a small proportion are calc-alkaline and may be related to subduction.

Structural history

With the exception of the sedimentary and metasedimentary rocks, the DFC is not intensely deformed. Deformation within the area has been concentrated in a marginal zone of high strain, the Cooper Bay dislocation zone (CBDZ) (see Section 10), and in narrow zones of high strain within the complex. The sedimentary and metasedimentary rocks were for the most part deformed prior to emplacement of the igneous rocks. However, the dip and strike of layering in the gabbro bodies in the Bogen Glacier area are similar in orientation to the banding of the paragneisses, and both are consistent with folding about F_3 fold axes. Since a new fabric and irregular folds are commonly developed within the magmatic granite-gneiss, the deformation may have been in part contemporaneous with the magmatic activity.

Narrow zones of high strain, mafic-felsic shear zones (Fig. 19) up to 2 m wide, deform the gabbro and mafic dyke suites. They are characterized by elongate white felsic veins within schistose mafic rocks. The zones, although variable in dip and strike, dominantly trend south-westerly and either dip gently north-west or steeply south-east. The change in orientation of mafic dykes within the shear zone indicates both dextral and sinistral movement within different shear zones.

Summary

The DFC records a complex interaction of mafic and felsic magmas with sedimentary and metasedimentary rocks through time. With the exception of Cooper Island, the southern part of the complex represents a deeper structural level where plutons cut complexly

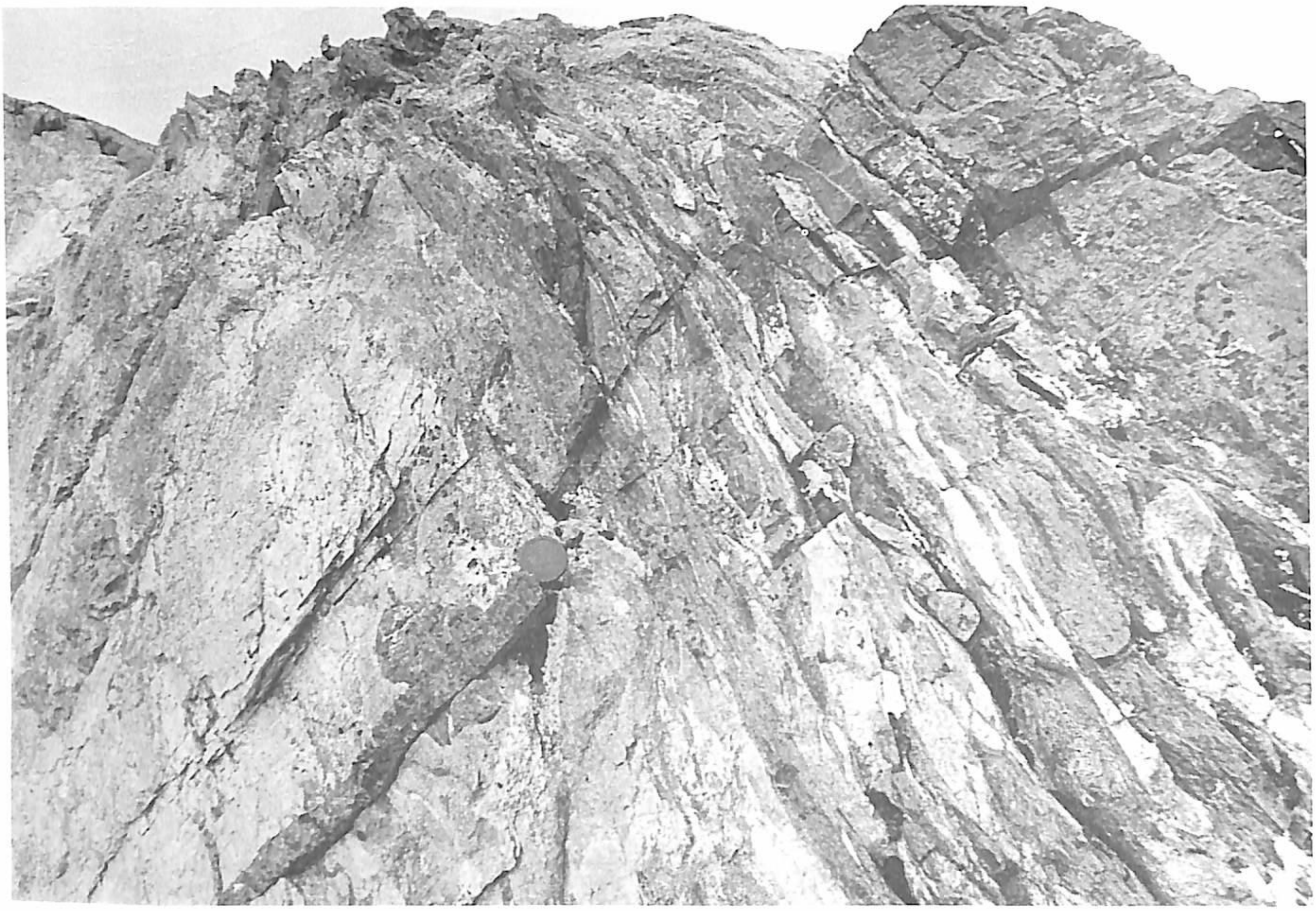


Fig. 19. Shear zone within a massive gabbro illustrating the deformation of a mafic dyke, Hamilton Bay. Lens cap is 5 cm in diameter.

deformed metasedimentary rocks. In its northern part and on Cooper Island, high-level magmas cut continentally derived clastic sedimentary rocks, on which they had a marked thermal effect. Although most of the deformation of the sedimentary rocks took place prior to formation of the igneous rocks, some was in part synchronous or post-dated their emplacement. The mafic and intermediate intrusions, dyke rocks and some of the granitic rocks formed by differentiation of a mantle-derived sub-alkaline tholeiitic magma during the Jurassic. Other granitic rocks formed by partial melting of the metasedimentary rocks.

4 Larsen Harbour Complex

Introduction

The south-western part of South Georgia is formed by a sequence of mafic and felsic lavas, estimated to be nearly 2 km thick. Associated with the lavas are pillow breccias, volcanogenic sediments, mafic and felsic dykes, a multiphase pluton (the Smaaland Cove intrusion) and minor gabbroic plutons (Bell and others, 1977; Storey and others, 1977; Storey and Mair, 1982; Mair, 1983, 1987).

The submarine volcanic rocks and sheeted dykes were termed the Larsen Harbour Formation (Bell and others, 1977). However, as the lavas, dykes and plutonic rocks are consanguineous parts of a single ophiolite sequence, it is suggested here that the term *Larsen Harbour Complex* (LHC) (first used by Dalziel, 1981) be used to describe the complete sequence. The term Larsen Harbour Formation (LH_1) should be retained for the lava sequence only, excluding the dykes.

The LHC covers a total area of about 120 km^2, underlying an area of rugged topography with sharp peaks and jagged ridges. The mountains are generally lower than those in the Drygalski Fjord Complex, with Mount Fraser (1611 m) the highest peak. Exposures extend from Cape Disappointment to Undine South Harbour and occur on Pillow Rock, at the easternmost end of Hauge Reef (Tanner and others, 1981). The following is a summary of the LHC, mostly after Mair (1987).

Stratigraphic and age relations

The LHC is separated from the DFC by a prominent topographic feature, interpreted as a fault zone, along its north-eastern margin. To the north it is terminated by an inferred fault along the line of Brøgger Glacier.

The age of the LHC is reasonably well constrained. It is conformably overlain by volcaniclastic sedimentary rocks of the Annenkov Island Formation (Tanner and others, 1981), part of the Mesozoic magmatic-arc sequence. The Annenkov Island Formation (AF) contains an Aptian–Albian fauna and deposition may have started as early as latest Jurassic times (see Section 5); it is intruded by andesitic sills and plutons, which have yielded ages of 100–103 Ma. A two point Rb–Sr whole-rock isochron obtained by Tanner and Rex (1979) for the Smaaland Cove intrusion gave an apparent age of 127 ± 4 Ma, with an initial $^{87}Sr/^{86}Sr$ ratio of 0.7061. This is interpreted either as the time of hydrothermal metamorphism soon after its emplacement or as the true age of crystallization, and it gives a minimum age for that part of the LHC. These data suggest that the LHC was formed during the latest Jurassic or earliest Cretaceous.

Although the LHC is not internally deformed, the lavas and tuffs are moderately inclined and the complex is deeply dissected by large fault zones. In the southern part the strata dip up to 54° towards the south-west, and in the central and northern part dips of up to 68° towards the west and north-west respectively are recorded. Despite the tilting and faulting, a pseudo-stratigraphy has been recognized (Fig. 20); the higher stratigraphic levels occur in the northern part and lower levels, intruded by the Smaaland Cove intrusion, in the southern part. Pillow lavas are widely developed at high levels in the sequence, massive lava flows occur throughout and amygdaloidal flows are restricted to low levels. Pillow breccias, volcanogenic sediments and cherts occur throughout the sequence, although tuffs and mudstones associated with the magmatic arc are more common in the higher levels. Mafic dykes occur at all levels, although they are rarer in the upper parts of the sequence; at low levels sections of up to 100 m of continuous mafic dykes are recorded. Lavas occur as limited screens between the dykes. The felsic dykes and lavas maintain complex relationships with the mafic rocks.

Mafic lavas and associated rocks

Dark grey or green amygdaloidal and vesicular pillowed lavas form about 60% of the extrusive rocks and are well exposed at Leon Head, Diaz Cove, Larsen Harbour and Pillow Rock (Fig. 21). The pillows are lensoid to sub-rounded in cross section but have complex shapes, commonly bulbous with a bifurcating form. They have a maximum size of 2.5 m long and 45 cm thick and occur in flow units up to 30 m thick. The chilled margins, which may be up to 4 cm thick are unornamented, rippled or intensely fractured.

Massive and amygdaloidal lavas form a minor part of the extrusive sequence but occur sporadically throughout. Massive lavas are abundant at Leon Head and amygdaloidal lavas at Rogged Bay and Nattriss Head. Columnar jointing and cooling cracks are locally developed at Leon Head, Diaz Cove and Pillow Rock (Fig. 22), and at Leon Head the tops of the flows often have a corrugated surface.

The lavas are fine-grained melanocratic porphyritic rocks with a variolitic or doleritic texture.

Lava breccias: The mafic lavas of the LHC are commonly associated with thick lava-breccia units (Fig. 23). They generally pick out interlava flow boundaries and may be

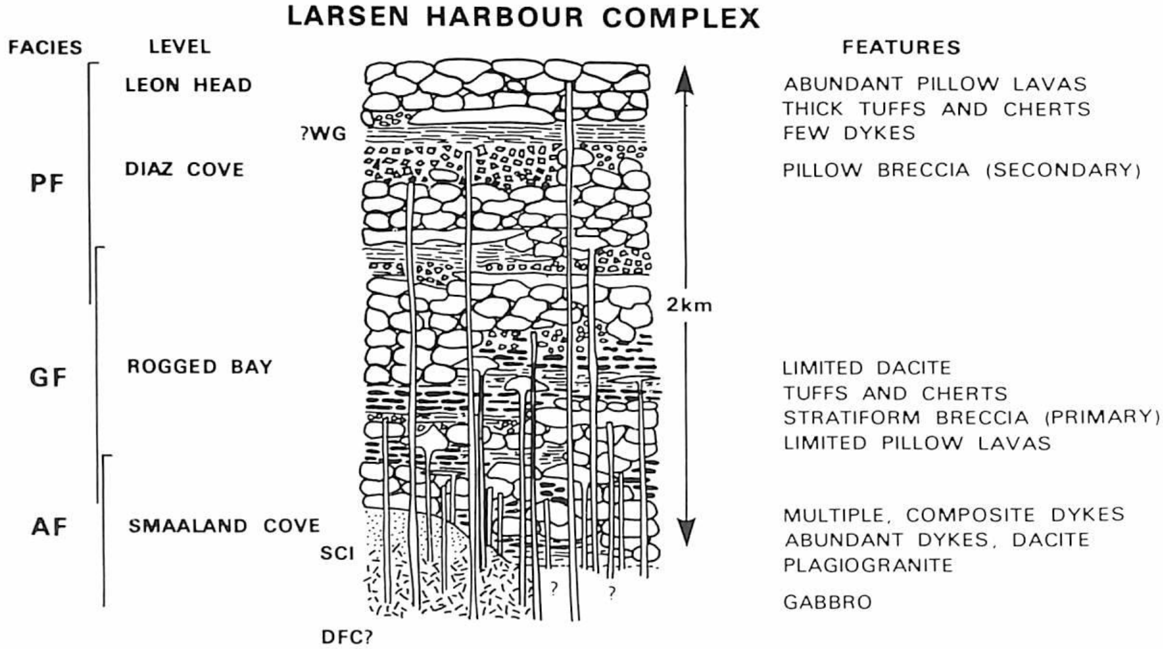


Fig. 20. Cartoon showing the main stratigraphic levels of the Larsen Harbour Complex, the areas in which the various levels are exposed and the distribution of the three major metamorphic facies (after Mair, 1987, fig. 32). PF = prehnite-pumpellyite facies; GF = greenschist facies; AF = amphibolite facies.

Fig. 21. A thick sequence of pillow lavas overlying massive amygdaloidal lava, Pillow Rock, Hauge Reef. Approximately 25 m of the section is shown.

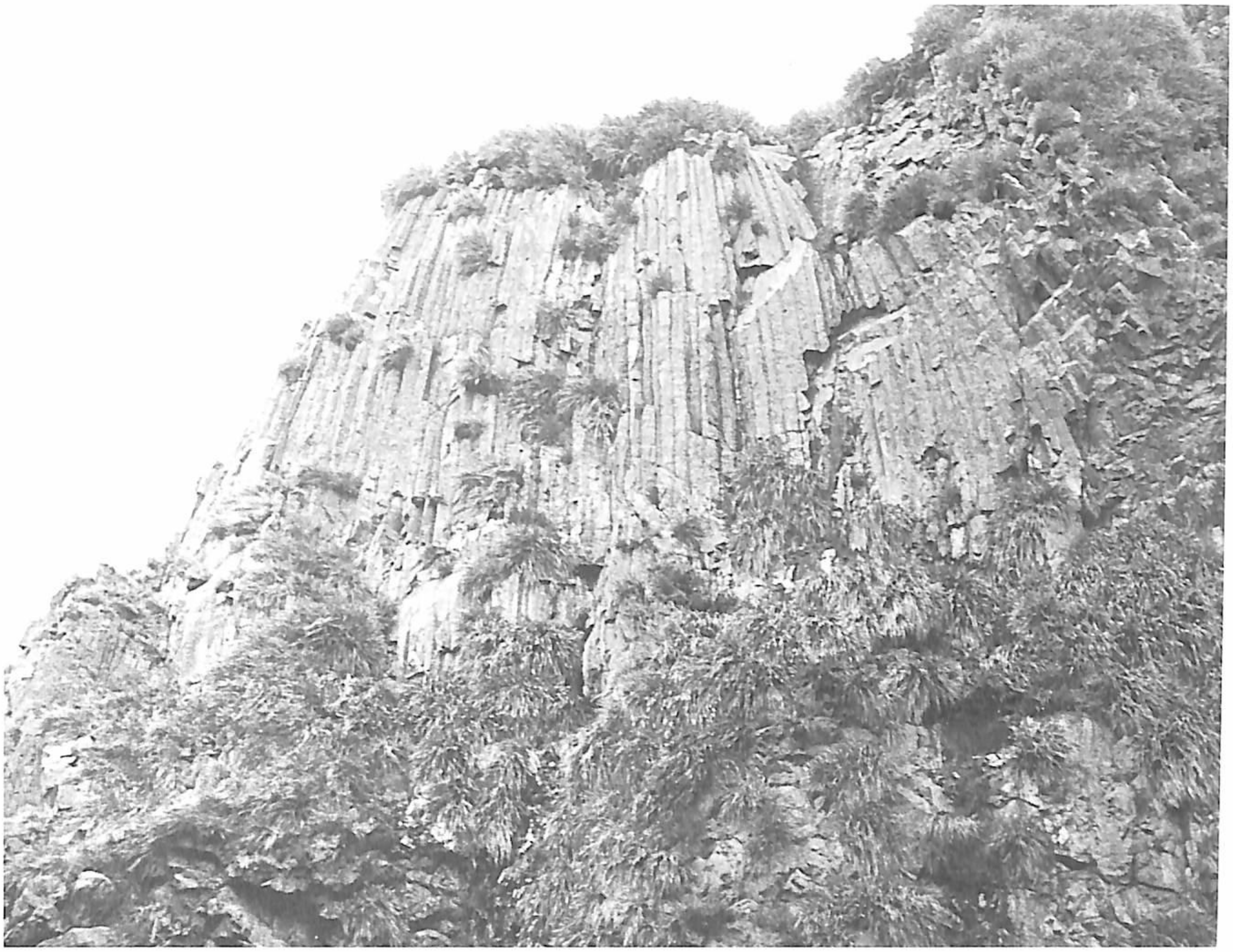


Fig. 22. Columnar-jointed basalt about 15 m thick overlying pillow lava, Pillow Rock, Hauge Reef.

associated with minor tuff and chert bands. Two main types have been recognized by Mair (1983, 1987).

Stratiform breccia is a bedded hyaloclastite deposit which is commonly concordant with tuffaceous sediments. It is composed of fresh, matrix-supported, rounded and elongate amygdaloidal basaltic lava globules or mini-pillows in an agglomeratic epidote, actinolite and chlorite-rich matrix.

Pillow-fragment breccia is of more local significance and lacks the bedded and well-sorted character of the stratiform breccia; at Leon Head for instance breccia infills a fissure developed within an earlier pillow lava flow. The fragments are highly angular, either clast- or matrix-supported, and include identifiable whole and fragmented pillows of varying textures, dacite and sedimentary rocks. The clast lithologies retain their original texture and lack the pervasive alteration rims of the fragments within the stratiform breccia.

The characteristics of the two types of breccia suggest they may have formed by different processes. The stratiform breccias possibly formed as a result of brecciation of hot magma during extrusion, perhaps the result of fountaining of submarine lava. They can be regarded as primary breccias since there was no intervening formation of pillow lavas. Transport of the hot hyaloclastite debris within a fluid medium would account for their bedded aspect. The pillow-fragment breccias were brecciated after consolidation by a secondary process and may have formed by a mechanical disintegration on the steep slopes of pillow volcanoes (Mair, 1987). Some may be talus breccias developed at the foot of submarine fault scarps and may have been deposited by debris flows.

Felsic rocks: Silicic lavas, dykes and small intrusions of dacite and rhyodacite form a minor but significant part of the LHC and crop out extensively in Doubtful Bay, Larsen Harbour, Smaaland Cove, Nattriss Head and Rogged Bay, implying that they are restricted to the lower stratigraphic levels of the mafic lava sequence. Local concentrations of both intrusive and extrusive types (up to 100 m wide by 50 m thick) may represent specific centres of acidic activity. At Nattriss Head a sequence of acidic rocks, up to 250 m thick, is unconformably overlain by bedded stratiform breccias. Contacts between the felsic rocks and enclosing mafic

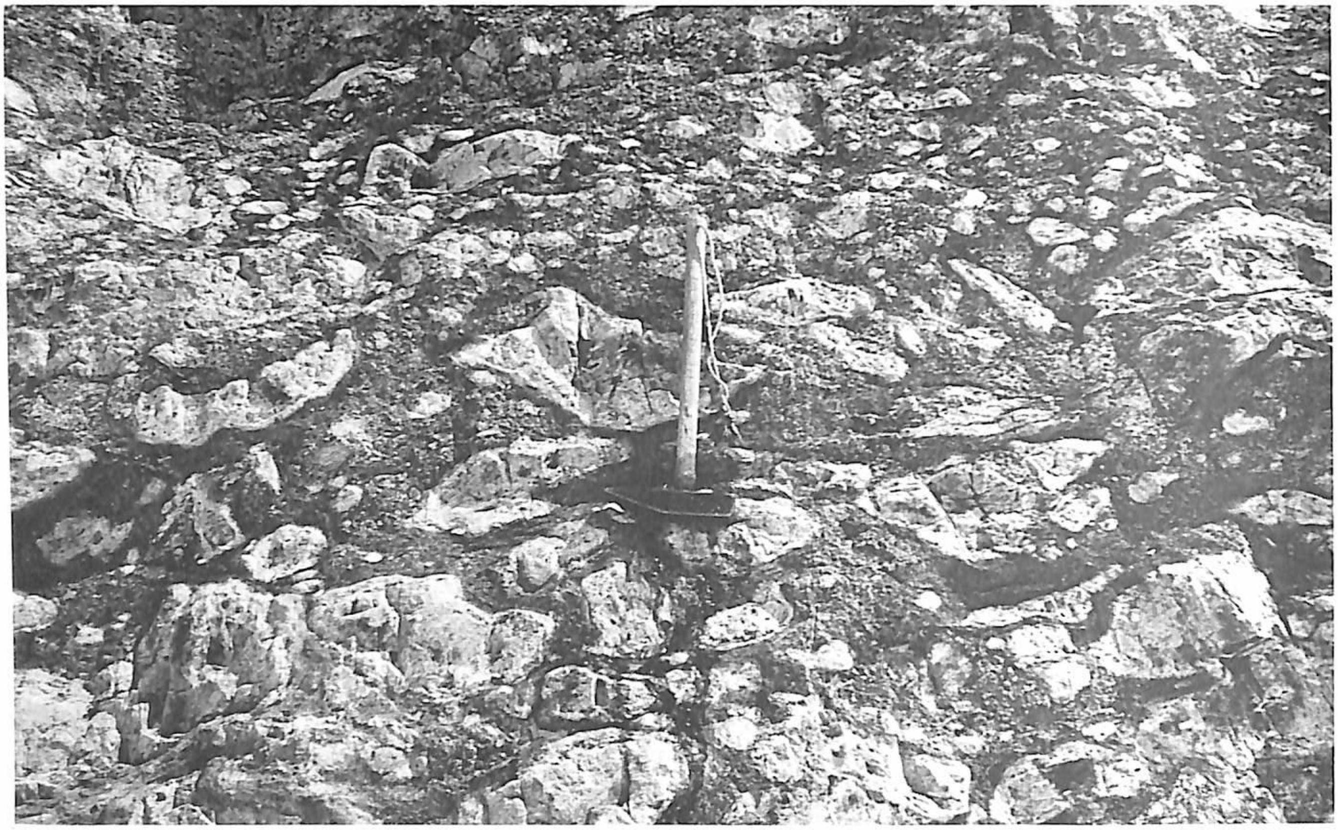


Fig. 23. Lava breccia within pillow lavas of the Larsen Harbour Complex, Smaaland Cove. The hammer shaft is 60 cm long.

Fig. 24. Volcanogenic sandstones and siltstones within the lavas of the Larsen Harbour Complex, Leon Head area. The scale is 10 cm long.

lavas are diffuse and commonly obscured by later mafic dykes. The felsic rocks were emplaced contemporaneously with the mafic lavas and pre-date the Smaaland Cove intrusion and some mafic dykes.

The extrusive lavas are generally pale-coloured amygdaloidal, porphyritic dacite and rhyodacite. Many have flow banding that is rarely regular and may be folded and locally steeply dipping. The basal and peripheral regions of the felsic lavas are commonly brecciated and consist of angular felsic fragments with rare mafic clasts. The fragmentation may be due to autobrecciation during extrusion and is a distinctive feature of the silicic rocks.

The silicic dykes and intrusions occur as truncated screens within the mafic dykes and are typically closely associated with the rhyolite domes. Unlike the mafic dykes, the felsic dykes have gradational and undulating chilled margins.

Volcanogenic sedimentary rocks: Lenticular units of volcanogenic sediment (Fig. 24) are interbedded at infrequent intervals within the mafic and felsic lavas and stratiform breccias. They range in thickness from a few centimetres to well-bedded sequences up to 10 m thick. Such well-bedded sequences occur near the head of Drygalski Fjord and at Larsen Harbour, Smaaland Cove and Leon Head. They infill undulations within lava-flow surfaces and occur as interstitial deposits between adjacent pillows. In general the stratification is not persistent but may continue along strike for up to 20 m.

The sediments consist of large-scale repetitive interbanded units, up to 2 m thick, of rhythmically interbedded and interlaminated green to dark green tuff and agglomerate, green, white and red cherts, and black mudstone. Complete sequences are rarely homogeneous and there is marked textural, colour and thickness variation. Sedimentary structures are well developed, especially in the thinner fine-grained units (< 7 cm), which show basal contacts with both load and flame structures. Although graded bedding is commonly well developed in the thinner beds, thicker beds commonly have a coarse base and fine-grained top with little intermediate grading. Most of the sedimentary rocks are derived from the breakdown of the LHC lavas but interbedded tuff and radiolarian-rich mudstone units, which are particularly common in the higher levels of the lava sequence at Leon Head and on Pillow Rock, may be derived from the island-arc sequence (Annenkov Island Formation).

Mafic dykes

Fine- to coarse-grained, porphyritic, single, multiple and composite mafic dykes of variable orientation form up to about 80% of the lower part of the LHC but are less common at higher levels. Although a sheeted complex (*sensu stricto*) has not been described, at Doubtful Bay adjacent mafic and rhyolite dykes form continuous exposures for up to 100 m (Mair, 1983). The single dykes are linear and straight sided, with distinct chilled margins. They are commonly surrounded by a series of quartz-filled veins which may be associated with faulting or shearing. The multiple dykes represent re-intrusion of magma along the axial zone of a pre-existing dyke, thus producing repetitive and identical sequences of chilled margins. Up to five intrusions of different grain size and texture have been recognized in the multiple dykes. The composite dykes show a gradation from a mafic margin to a more felsic interior and represent a change in magma composition with time. The dykes show complex orientation patterns and five major trends have been recognized by Mair (1983). In the south-east dykes generally trend 075° or 132°, whereas 022°, 132°, and 174° trends are common in the northern part of the complex. Most dykes dip eastwards at high angles although a distinctive group of dykes at Nattriss Head, Ranvik, Kupriyanov Islands and Diaz Cove dip at 30° or less. The age relationships between the different dyke trends are not consistent although the 132° trend is commonly the earliest.

The dyke rocks show considerable variation in grain size, texture and plagioclase composition and are predominantly medium- to coarse-grained, porphyritic dark grey melanocratic rocks with ophitic, intersertal and basaltic textures.

Smaaland Cove intrusion and minor isolated gabbro bodies

At Smaaland Cove, on the southern coast of the island, a multi-phase pluton, about 6 km² in area, intrudes mafic lavas of the lower levels of the LHC (Fig. 25). It cuts the older and is intruded by the younger dyke suites. On its eastern side, it intrudes and brecciates the overlying volcanic rocks, and xenoliths and screens of partly assimilated lava are common; elsewhere the upper contact is sub-horizontal and sharp with limited diffusion or interaction of material. Its eastern contact is terminated abruptly, possibly by a fault along Doubtful Bay.

The Smaaland Cove intrusion varies in composition from quartz-gabbro to alkali granite (Fig. 26). It was tentatively described as a quartz-diorite (Bell and others, 1977), later classified as a tonalite (Storey and others, 1977) but is now referred to as the Smaaland Cove intrusion following detailed mineralogy by Mair (1987). Small-scale variations in grain size and colour are a common feature but no distinct boundaries between different phases are evident.

Minor gabbros: On the north side of Wheeler Glacier a coarse-grained gabbro less than 1 km² intrudes the lava sequence and there is also an isolated gabbro body at the head of Novosilski Glacier.

Metamorphism

The primary mineralogy of the mafic and felsic, volcanic and plutonic rocks of the LHC is extensively replaced by secondary assemblages (see Mair, 1987). In many cases evidence of the primary phases is only preserved as small relicts within complex pseudomorphs. Secondary phases define a metamorphic zonation (established by microprobe analysis) from lower amphibolite facies in lower levels of the LHC, to greenschist and prehnite-pumpellyite facies in higher levels. The dykes contain assemblages of the above three facies and the underlying plutonic rocks contain amphibolite-facies assem-

Fig. 25. Prominent roof contact between the pale granitoids of the Smaaland Cove intrusion and dark pillow lavas; west side of Smaaland Cove. The contact is cut by later basic dykes. The mountains in the background are approximately 500 m above sea level.

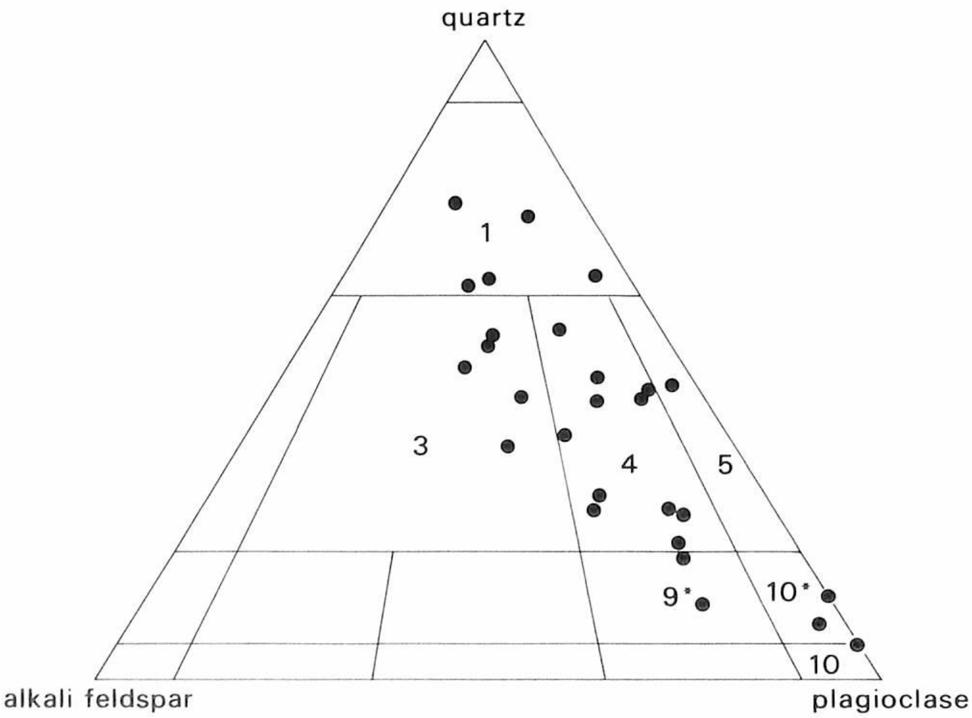


Fig. 26. Composition of the Smaaland Cove intrusion. Classification after Streckeisen (1973); 1, quartz-rich granitoids; 3, granite; 4, granodiorite; 5, tonalite; 9*, quartz-monzodiorite, quartz-monzogabbro; 10, gabbro, diorite; 10*, quartz-diorite, quartz-gabbro.

blages locally retrogressed to greenschist facies.

Although the metamorphism broadly defines a zonation, retrogressive and disequilibrium assemblages are common and the boundaries between adjacent metamorphic facies are complex and irregular. This is typical of a hydrothermal system within oceanic rocks (Coleman, 1977) and the sporadic alteration within the low-level plutonic rocks and dykes indicates a limited depth of fluid movement. Although the metamorphic zonation has probably been caused solely by a hydrothermal system, there is a possible burial metamorphic imprint (Mair, 1987).

Geochemistry

Major- and trace-element analyses of 28 mafic lavas, 24 mafic dykes, 17 felsic dykes and 21 plutonic rocks from the LHC have been carried out (for full details see Mair, 1987). The mafic rocks are predominantly olivine-tholeiites with some quartz-normative tholeiites and, together with the felsic rocks, including the dykes, lavas and intrusive rocks, form part of a bimodal comagmatic mafic-felsic suite produced by differentiation of a subalkaline tholeiitic magma. They have a similar chemistry to the mafic and some of the felsic rocks in the DFC (see Figs. 16–18), were derived from the same magmatic source and were emplaced in the same back-arc basin tectonic setting.

Compared to mid-ocean ridge basaltic suites (MORB), the mafic rocks of the LHC have similar high field strength element abundances (Ti, Y, Zr, Nb and P) and plot within the mid-ocean ridge basalt field on the Ti-Zr-Y and Ti-Zr discrimination diagrams of Pearce and Cann (1973). They are however enriched in the large ion lithophile elements (Sr, K, Rb, Ba, La and Ce), a characteristic of mafic rocks emplaced in a back-arc-basin setting which may be related to the influence of the subducting slab (Saunders and Tarney, 1984).

The felsic rocks have chemical characteristics typical of oceanic plagiogranites formed by differentiation of an ocean-floor basalt (Coleman and Peterman, 1975). They are enriched in Na_2O and have both high and low K_2O values. As the K_2O-rich rocks are enriched in Zr and have equivalent Ce/Y ratios to the K_2O-poor rocks, it is concluded that they were derived by differentiation of the basic magma and that they may have been contaminated by partial melting of lenses of metasediment trapped within the mafic suite.

5 Annenkov Island Formation

Introduction

The Annenkov Island Formation (AF) was originally described as a facies variant of the Cumberland Bay Formation (Trendall, 1959) but Pettigrew (1975, 1981) defined it as a separate formation, with type sections on Annenkov Island. Tanner and others (1981) extended Pettigrew's definition to include rocks on the Pickersgill Islands, the western and central islands of Hauge Reef, Low Reef and Mislaid Rock. The whole formation is 3–4 km thick and consists of two members: a 2–3 km thick Lower Tuff Member (AF_1) and an Upper Breccia Member (AF_2), at least 1.2 km thick. Both members are the volcaniclastic products of contemporaneous andesitic volcanism, both are metamorphosed to zeolite facies and both have been intruded by various small hypabyssal bodies. The AF is gently tilted to the west at 10°–30° and is cut by near-vertical normal faults.

Stratigraphic and age relations

The upper parts of the Larsen Harbour Formation (LH_1) exposed at Leon Head (Mair, 1987) and on Pillow Rock (Tanner and others, 1981) contain significant amounts of andesitic tuffs interbedded with the pillow lavas. This, coupled with the overall gentle westerly dip of both the LH_1 and AF led Tanner and others (1981) to suggest a conformable relationship between the two formations. Sediments from Pillow Rock have yielded spores and acritarchs which indicate a latest Jurassic or earliest Cretaceous age (personal communication from Dr N.F. Hughes).

There is a fauna of ammonites and bivalves in the upper parts of the AF_1 on Annenkov Island (Wilckens, 1932, 1947; Trendall, 1959; Thomson and others, 1982). Many of the constituent species appear to be endemic to Annenkov Island and can only be assigned a broad early Cretaceous age (Thomson and others, 1982). The AF_1 has also yielded foraminifera, cirripedes, fish scales and bones, a large vertebrate fragment, cycad fronds and fossil wood. The contact with the overlying AF_2 appears conformable (Pettigrew, 1981) but the 15°–20° divergence between the attitude of the base of the AF_2 and bedding in the AF_1 led Storey and Macdonald (1984) to suggest that the boundary was the erosive base of a major channel. The AF_2 has yielded fragments of the belemnite *Dimitobelus*, suggesting an Aptian–Albian age (Pettigrew and Willey, 1975). The top of the AF_2 is not seen, but the highest known levels crop out on Mislaid Rock.

The Annenkov Island Formation is intruded by irregular sheets and stocks of andesite and diorite. Hornblende from these intrusions yielded K–Ar mineral ages of 85–103 Ma and a monzodiorite intrusion from the Pickersgill Islands gave a whole-rock Rb–Sr isochron age of 81 ± 10 Ma (Tanner and Rex, 1979). Of special interest are the 100–103 Ma mineral ages from intrusions in the AF_2. Pettigrew (1975, 1981) concluded that these were intruded into unconsolidated sediment and the coincidence of these mineral ages with the late Aptian–Albian age (100–110 Ma) for the AF_2 suggested by Pettigrew and Willey (1975) supports this conclusion.

Sedimentology

Lower Tuff Member: Dark grey, well-indurated radiolarian and tuffaceous mudstones, interbedded with tuff beds and lenses of radiolarite form the AF_1 (Fig. 27). The mudstones can be structureless or show faint parallel or cross lamination and include concentrations of recrystallized radiolaria which may be lag deposits.

The proportion of tuff beds within the AF_1 varies from 30–40% in the lower part of the section to less than 10% at higher stratigraphic levels (Pettigrew, 1981), although tuff may be locally dominant (60–70%; Storey and Macdonald, 1984). The beds are generally 3–10 cm thick, though beds of up to 40 cm are relatively common, and they may reach 2 m. Beds are sharp based, occasionally erosive and in rare instances show linear tool marks. Most show good distribution grading and many contain mudstone intraclasts. Parallel lamination is common and ripple cross lamination is also found.

Pettigrew (1981) thought that some of the beds could be turbidites, but that most were due to airfall. However, Storey and Macdonald (1984) concluded that turbidites were dominant over airfall and they recognized T_{ae}, T_{abe}, T_{be}, T_{ace} and T_{ce} types. Slump structures, synsedimentary faults, centimetre-scale sedimentary dykes and load casting are all common. These structures point to deposition of saturated sediment in an unstable environment, probably on a slope.

Basal sandstone: Pettigrew (1981) defined the basal sandstone unit as part of the Upper Breccia Member. It is, however, sufficiently distinct to warrant separate discussion. The basal sandstone unit (Fig. 28), found only on Annenkov Island, is 91 m thick, and comprises very coarse sandstone and granulestone with subordinate pebbly granulestone. Beds are 0.5–2 m thick, with sharp, flat or slightly undulating bases and sharp flat tops. They are either amalgamated, or separated by very thin mudstone layers (1–10 cm). Most beds are structureless

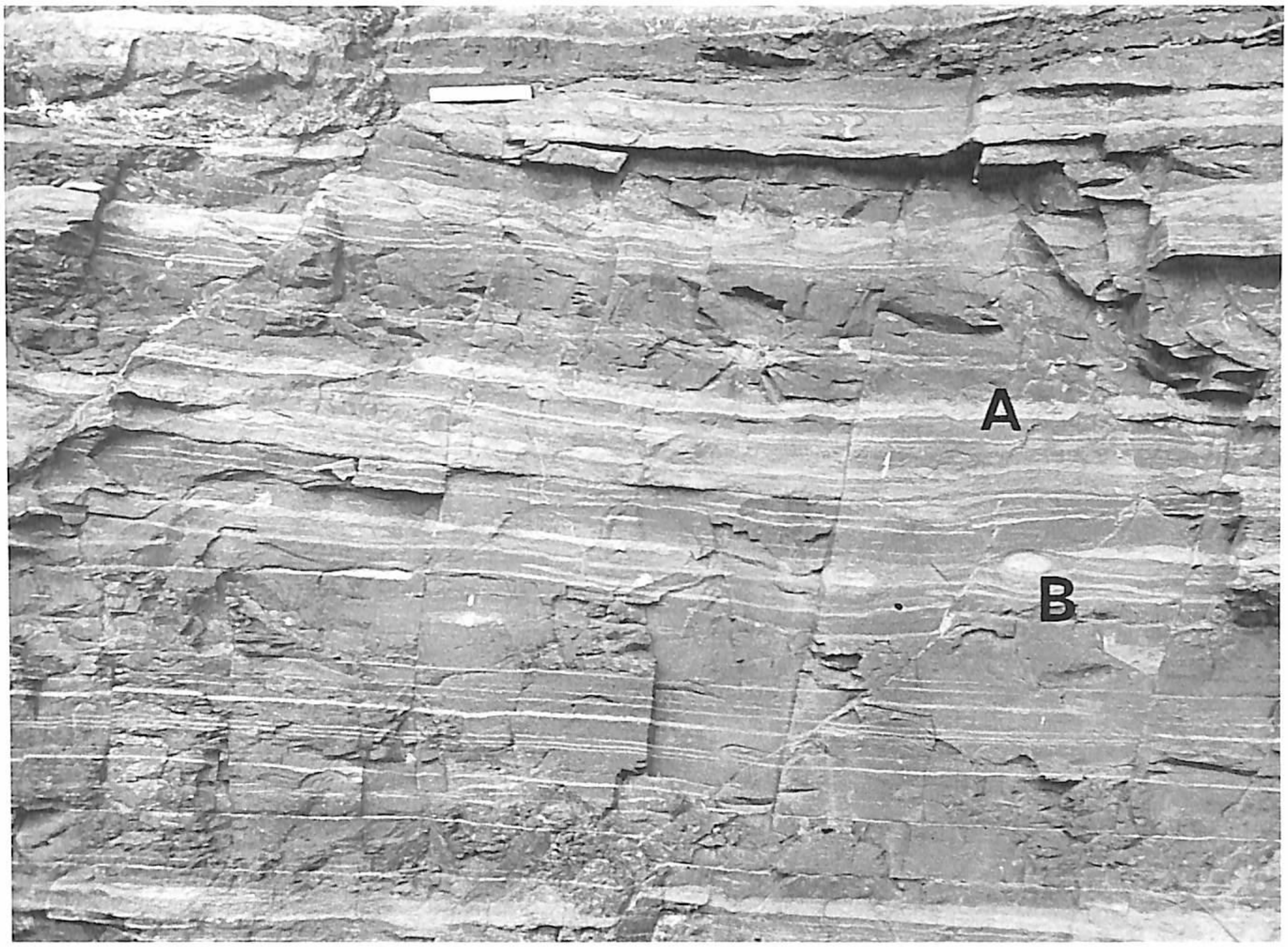


Fig. 27. Thin-bedded tuffaceous rocks of the Lower Tuff Member on the westernmost island of Hauge Reef. A, graded crystal-tuff showing synsedimentary faulting to left of centre; B, calcareous concretion. The scale is 15 cm long.

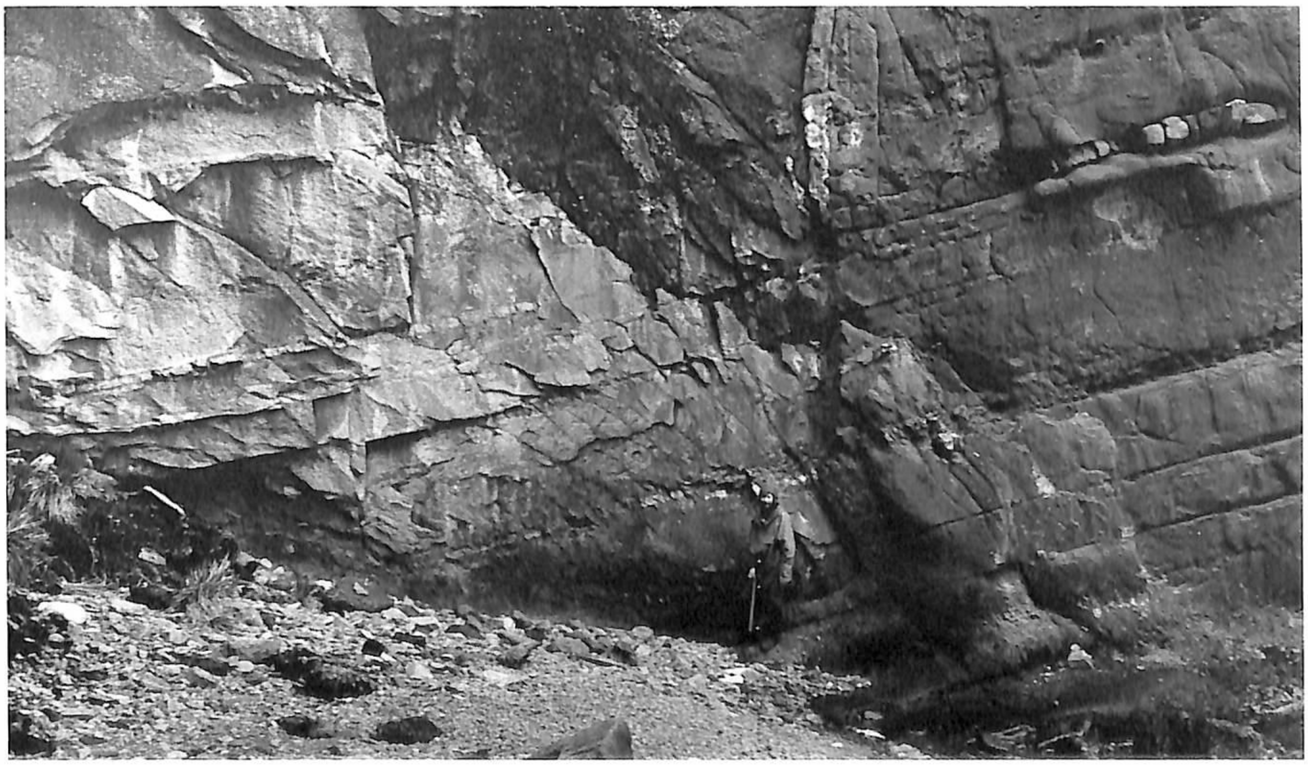


Fig. 28. Basal sandstone of the Upper Breccia Member, north coast of Annenkov Island.

Fig. 29. Crudely bedded sandstone lens within conglomeratic breccias of the Upper Breccia Member, Low Reef. The hammer handle is 60 cm long.

or show faint parallel lamination, especially in their upper parts, and contain mudstone intraclasts, which reach 30 cm in some cases. The basal sandstone is overlain, apparently conformably, by the main sequence of breccia.

Upper Breccia Member: The AF_2 (Fig. 29) is andesitic, and clasts closely resemble the intrusive andesites (Pettigrew, 1981). The breccias can be matrix or clast supported; they are crudely bedded on a 1–5 m scale, and intercalated with subordinate irregular beds of very coarse sandstone or (more rarely) mudstone and tuff of the AF_1 facies. Bed bases are commonly erosive whereas tops are generally flat, or have protruding clasts which are draped by overlying sediment (Storey and Macdonald, 1984). Most of the breccias appear structureless or display an irregular horizontal shearing lamination. Bedded breccias usually pass both horizontally and vertically into very thick sequences of massive disorganized breccia (Pettigrew, 1981). The evidence of basal erosion in the bedded portions of the sequence suggests that these may be the result of bed amalgamation. Drag folds below bed bases, and large intraclasts of tuff and mudstone in both massive and bedded breccias (Pettigrew, 1981) offer further support to this theory.

Depositional processes: The AF_1 is a mixed unit of suspension deposits (mudstones), volcanic airfall and volcaniclastic turbidites, with probable reworking by oceanic bottom currents (Storey and Macdonald, 1984). The lack of wave-generated structures and the presence of slumps suggest that deposition took place below wave base, perhaps on the outer shelf or upper slope.

The sandy unit at the base of the AF_2 represents the deposits of high-density turbidity currents (usage of Lowe, 1982). The overlying breccias were deposited from submarine debris flows (Pettigrew, 1981; Storey and Macdonald, 1984). Pettigrew (1981) thought that the overall upward coarsening in the Annenkov Island Formation represented progadation of the steep flanks of a marine andesitic volcano, but Storey and Macdonald (1984) thought it more likely that the AF_2 was deposited in a large-scale channel incised into the AF_1.

Composition: The whole formation is andesitic in composition. Tuffs are almost all crystal-lithic with a variable proportion of vitric material (0–20%) and matrix commonly forms 30–50% of the rock. The lithic clasts are mainly andesites of various types with plagioclase the dominant crystal type. There are rare laumontitized vitric tuffs on Annenkov Island (Pettigrew, 1975), and Tanner and others (1981) found possible lappilli-tuffs on Hauge Reef. The sandstones interbedded within the AF_2 and forming the matrix to the breccias are lithic greywackes and feldspathic greywackes (usage of Pettijohn and others, 1972), comprising andesitic fragments with subordinate dacite and microdiorite in a zeolitized mudstone matrix.

The petrography suggests derivation of all the rocks from a contemporaneous calc-alkaline volcanic arc. This theory is supported by geochemical analyses of tuffs,

Fig. 30. Irregular stocks of pale grey diorite cutting darker mudstone and tuff of the Lower Tuff Member on the central island of Hauge Reef. The cliff is about 25 m high.

mudstones, breccias and associated plutonic and volcanic rocks from the AF (Storey and Tanner, 1982).

Igneous intrusions

The igneous rocks intruded into the AF span the range from gabbro to granite, with most being relatively early stocks and sheets and later dykes of andesite (Pettigrew, 1981; Tanner and others, 1981). The early intrusions are dominantly hornblende-andesites with subsidiary biotite, whereas later andesites are porphyritic biotite-andesites.

On Annenkov Island the hornblende-andesites form intrusions with irregular margins in the AF_2 and sill-like bodies near the contact between the AF_1 and AF_2. They occur as steep-margined irregular stocks on islands along Hauge Reef and in the Pickersgill Islands (Fig. 30), commonly with chilled margins against the sedimentary rocks but rarely displaying a contact aureole. Sills of hornblende-andesite are present in the Pickersgill Islands, where there is also a large irregular andesite pluton with sill-like marginal apophyses. On the westernmost island of Hauge Reef there is a large tabular body of biotite-andesite, which is possibly later than the hornblende-andesite; biotite-andesite also forms some of the stocks on the central island of the reef. The later andesites form parallel-sided dykes up to 2 m wide on Tanner Island and several of the smaller islands, where they cross-cut almost all of the other intrusions.

There are also larger intermediate intrusions in the Pickersgill Islands. All of the central island of the group is made up of a hornblende-microdiorite body of unknown form and Tanner Island is a composite intrusion, mainly of quartz-monzodiorite. The Tanner Island intrusion (81 ± 10 Ma) shows evidence of an early basic phase followed by paler, more acidic rocks.

The basic rocks of the area form early sills, late dykes and a small pluton. Sills of metabasite and spilite up to 5 m thick occur in the AF_1 of Annenkov Island and Hauge Reef. These can be transgressive and have pillowed, sediment-injected margins indicative of intrusion into wet sediment. They are the earliest of all intrusions. On Tanner Island there are basaltic dykes, 2–3 m thick, which cut almost all the intrusions including the late andesite dykes mentioned above. A larger basic intrusion forms much of the northern island of the Pickersgill Islands. It is a varied body of hornblende-pyroxene-gabbro, leuco- and meladiorite and quartz-diorite; the leucodiorite and quartz-diorite form the margins of the intrusion.

The latest intrusions of all are microgranite dykes on some of the Pickersgill Islands. These are almost entirely quartzo-feldspathic, apart from subsidiary chloritized biotite and on Tanner Island they are associated with aplite sheets.

Storey and Tanner (1982) analysed a representative suite of igneous and sedimentary rocks from the AF. The analyses define a strong calc-alkaline trend and most fall into the calc-alkaline basalt fields of various discriminant plots. They interpreted all the sediments and the vast majority of the intrusions as part of a coeval calc-alkaline island-arc complex and pointed out the possibility of contamination by continental crust, as indicated by high K_2O, Rb, Sr and Ba values. The only exceptions to this are the spilite and metabasite sills which fall within the marginal-basin basalt field on a Ce/Y plot, and in the ocean floor/MORB fields of all the triangular discrimination plots proposed by Pearce and Cann (1973). Tanner and others (1981) had suggested on the basis of petrography and field relations that the basic sills were derived from the same source as the pillow lavas of the Larsen Harbour Formation. From the geochemical evidence, Storey and Tanner (1982) supported this conclusion and suggested that there was overlap between the calc-alkaline volcanic activity of the magmatic arc and the tholeiitic activity represented by the Larsen Harbour Formation.

6 Cumberland Bay Formation

Introduction

The Cumberland Bay Formation (CB) forms well over half the island and most of the high peaks. It is a thick sequence of interbedded volcaniclastic greywackes and shales, differentiated from the superficially similar quartzose Sandebugten Formation (SF) on petrographic grounds.

At various times the CB has been subdivided into "members" (Ferguson, 1915) or been included with the Sandebugten Formation as a single series (Wordie, 1921; Holtedahl, 1929). Ferguson (1915) recognized that the rocks in the Cape George area were different from most of the outcrop, and proposed a differentiation between the "Cape George Series" and a tripartite "Cumberland Bay Series". Tyrrell (1930) proposed a unitary "Cumberland Bay Series" unconformable on a quartzose "Godthul Harbour Series". This scheme was the basis for most subsequent stratigraphies except that the quartzose rocks were renamed the "Sandebugten Series" (Trendall, 1953). The stratigraphy was formalized by Stone (1975, 1980) who proposed the names Cumberland Bay and Sandebugten formations and differentiated an outlier of the CB as the Barff Point Member.

Estimates for the maximum thickness of the formation vary from 10 km (Trendall, 1959) to 5–8 km (Tanner and Macdonald, 1982). Calculation is hampered by the complete absence of stratigraphic markers within the CB and the pervasive deformation by large-scale overturned chevron folds which face and verge north-east and plunge gently towards the north-west. This is also a severe limitation to detailed sedimentological research (Tanner and Macdonald, 1982; Macdonald, 1986).

Stratigraphic and age relations

There are no depositional boundaries between the CB and any other unit on South Georgia. Along the south-west side of the outcrop the CB is separated from the Drygalski Fjord Complex by the Cooper Bay dislocation, which is inferred to continue up the lower Spenceley Glacier and across Undine South Harbour, separating the CB from the Larsen Harbour Complex (Storey, 1983*a*), and the Ducloz Head Formation (Storey, 1983*b*).

On the east side of the outcrop the CB overlies the Sandebugten Formation along a westerly dipping thrust plane which is exposed on Dartmouth Point, the eastern spurs of Mount Brooker, Will Point and Cape Charlotte (Stone, 1980). Trendall (1953) postulated that the contact was tectonic and showed that there was a contact on Dartmouth Point separating rocks with different vergence and facing directions (Trendall, 1959). The contact was actually located by Dalziel and others (1975), who demonstrated that it was a thrust with a 1–2 cm thick mylonite zone. Stone (1975, 1980) examined the thrust on Will Point and at various localities between Royal Bay and Gold Harbour and showed that all were similar to the Dartmouth Point locality, with the CB rocks being overthrust towards the north-east. Mair (1981) found that on Dartmouth Point the thrust plane has a complex outcrop pattern due to later normal faulting. Recently Craw and Turnbull (1986) have refined the position of the thrust plane in the area east of Mount Brooker.

Stone (1975, 1980) found an easterly dipping thrust south of Barff Point, with CB rocks overlying it, confirming the earlier report of Aitkenhead and Nelson (1962). In contrast, Dalziel and others (1975) considered that all of the rocks in the area belonged to the Sandebugten Formation. Tanner (1982*b*) re-mapped the area, corroborated Stone's findings and stated that "In most respects the thrust contact on the Barff Peninsula is the mirror image of that on Dartmouth Point . . ." (Tanner, 1982*b*, p. 173). The outlier of CB to the north-east of this thrust is designated the Barff Point Member, following Stone (1980).

The inferred southern boundary of the CB, against the Cooper Bay Formation, extends along Twitcher Glacier (Storey, 1983*a*). The nature of the contact is unknown but Tanner and others (1981) suggested that it may be a continuation of the main thrust fault. This implies a single thrust plane, separating quartzose formations (Sandebugten and Cooper Bay formations) from the volcaniclastic CB, deformed in a broad anticline–syncline pair plunging north-west.

The CB is almost unfossiliferous but Thomson and others (1982) reviewed previous palaeontological work and reported a number of new fossil localities. They concluded, mainly on the basis of sparse occurrences of the bivalve *Aucellina* and a few fragments of large heteromorph ammonites, that most of the CB was Aptian–Albian in age. Belemnite fragments from erratic blocks of CB from the Royal Bay area provided the only evidence for pre-Aptian strata: one of these was thought to have affinities with *Belemnopsis*, suggesting a late Jurassic or Neocomian age (Stone and Willey, 1973). However recent work by Doyle (1985) discounts the presence of *Belemnopsis* and points only to a Mesozoic age for the collection described by Stone and Willey.

Thomson and others (1982) also carried out radiometric age dating on 12 slate samples from the CB, using K–Ar whole-rock methods. Their results range from 51

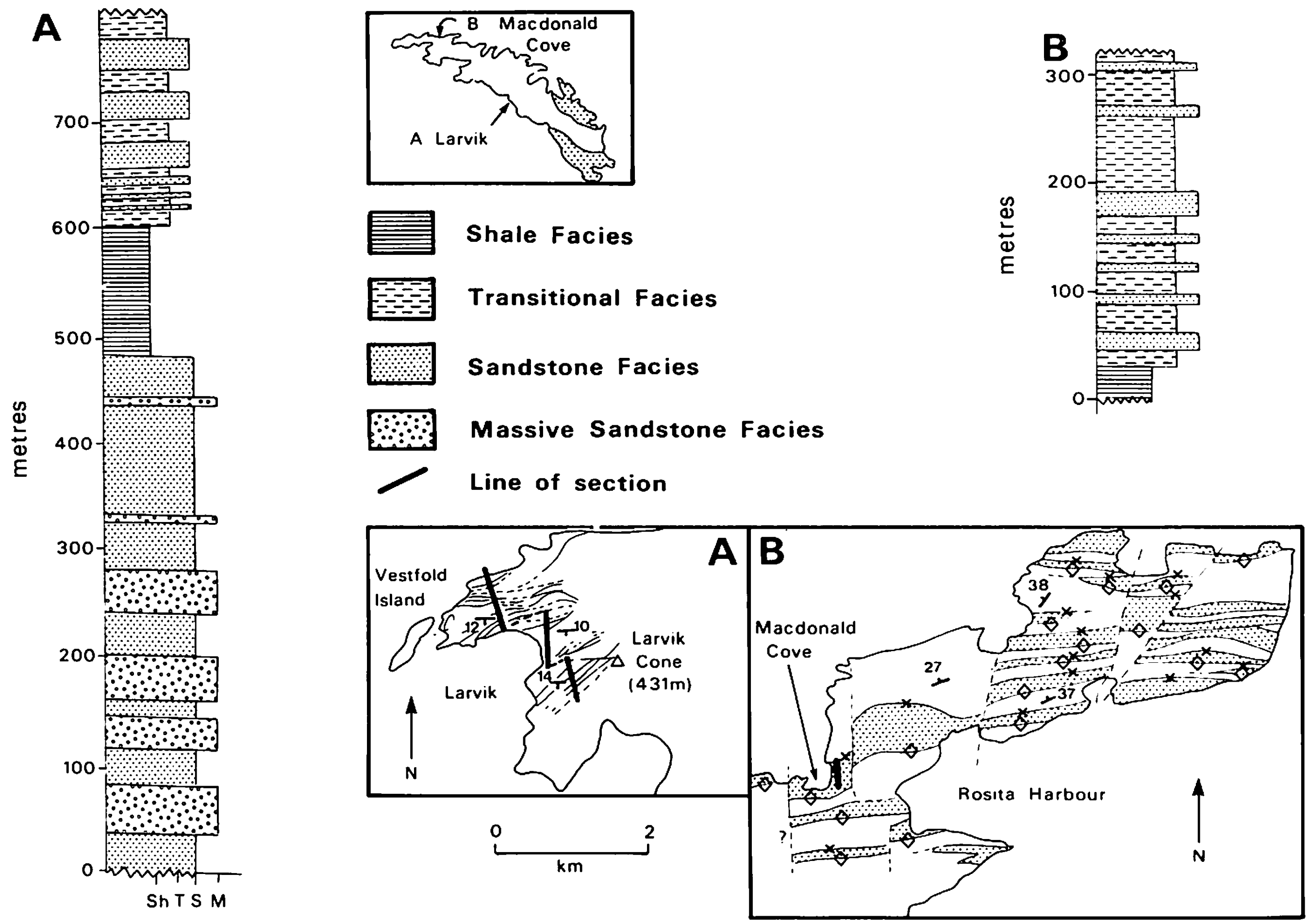


Fig. 31. Measured sections at the two proposed type localities of the Cumberland Bay Formation: A, Larvik; B, Macdonald Cove. The insets show the location of the sections in relation to major unit boundaries at Larvik (A), and to principal folds in the Macdonald Cove area (B). On inset B inverted limbs are stippled.

± 2 to 135 ± 5 Ma (recalculated to 51–138 Ma using current conventional isotopic constants, *cf.* Steiger and Jäger, 1977), with all the older dates coming from the south-western coastal exposures, where specimens contained flakes of detrital white mica. Discounting specimens with detrital mica led to a much tighter distribution of ages, with five samples giving 84–93 Ma. These dates were interpreted as uplift and cooling ages.

The evidence suggests that the bulk of the CB was deposited during the Aptian–Albian (119–97 Ma), over a period of *c.* 20 million years. However, evidence from the Annenkov Island Formation that arc activity started in the latest Jurassic, suggests that deposition of the CB may have begun as early as 144 Ma ago, extending the time of deposition to 50 million years. Deformation of the CB probably took place in the mid-Cretaceous, with the main period of uplift at 84–93 Ma (Cenomanian–Santonian) although uplift may have continued until the early Eocene. The concordance of the uplift ages of the CB with the age of plutonism in the AF is striking.

Type sections

When Stone (1980) revised the stratigraphic nomenclature of South Georgia and proposed the *Cumberland Bay Formation* as a formal name, he did not define a type section. He hoped that at some future time a type section would be found in the Cumberland Bay area. Further work has shown that there are no suitable sections around Cumberland Bay, where repeated chevron folding and zones of second-phase deformation result in there being no long undeformed sections (Tanner and Macdonald, 1982). The variability of the CB (Macdonald, 1986; also discussed in the next section) means that a single type section cannot adequately encompass the variety of bed types present.

Two type sections are proposed:

(a) *Larvik:* An 800 m section has been measured from the northern side of the point adjacent to Vestfold Island to the north face of Larvik Cone (Figs. 31 and 32). This section is typical of the outcrop on the south-west coast in having a high proportion of massive sandstone facies, but it also displays a good range of the finer-grained lithologies.

(b) *Macdonald Cove:* A 320 m section is exposed on the eastern side of the cove on the common limb of a large anticline–syncline pair (Figs. 31 and 33). Although strata are inverted, exposure is excellent, a wide variety of lithologies is present and the section contains one of the few fossil localities found in the CB.

Sedimentology

The foundations of all later sedimentological research

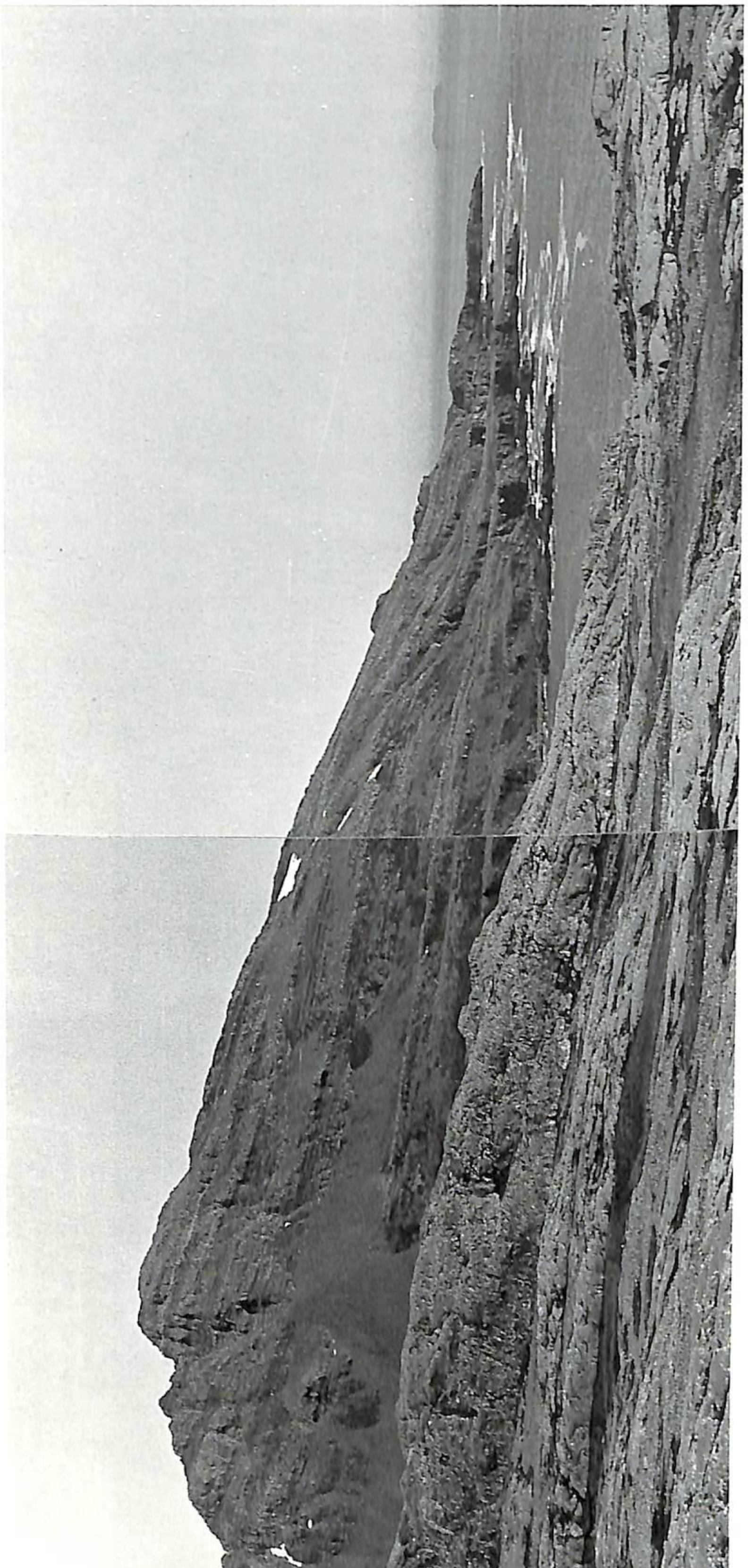


Fig. 32. View south across the Larvik area looking from base to top of one of the type sections of the Cumberland Bay Formation.

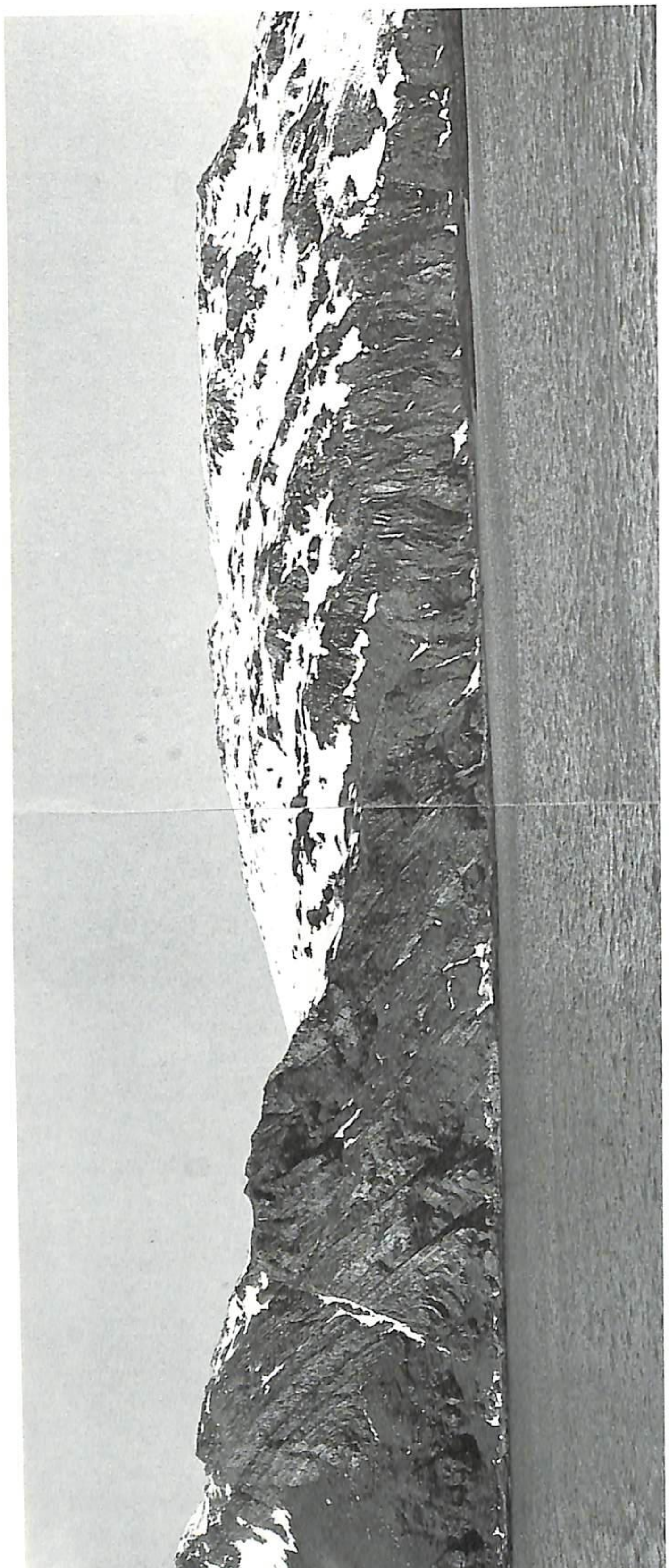


Fig. 33. View east across Macdonald Cove showing one of the type sections of the Cumberland Bay Formation. The section extends from the prominent gully just right of centre to the point where the axial plane of the major syncline cuts the shoreline. Note that strata are inverted.

Fig. 34. Turbidite bed (T_{abcde}) from the Cumberland Bay Formation near Prince Olav Harbour. The sandstone bed is 55 cm thick.

were laid by Trendall (1959), who recognized that most of the sandstone beds were turbidites (Fig. 34) and suggested, on the strength of cross laminae in the upper parts of the beds, that flow was towards the north-west. Later research confirmed Trendall's idea of north-westerly flow (Dott, 1974; Dalziel and others, 1975; Stone 1975, 1980; Winn, 1978; Tanner, 1982*b*) although some authors obtained inconclusive or contradictory results from small numbers of readings (Frakes, 1966; Skidmore, 1972; Mortimore, 1979). Dott (1974) and Tanner (1982*b*) emphasized the need for accurate structural control on the untilting of palaeocurrent indicators. Little sedimentological research other than palaeocurrent studies and petrography was carried out prior to the 1975–76 season, although both Stone (1975, 1980) and Mortimore (1979) recognized the existence of at least two facies of the CB. Fuller reviews of early sedimentological research can be found in Tanner and Macdonald (1982) and Macdonald and Tanner (1983).

Facies: The CB consists of alternations of beds of greenish-grey volcaniclastic lithic greywacke and dark grey shale. Most of the greywackes are medium–coarse sandstone grade, although very coarse sandstone and granulestone is relatively common on the south-west coast. Stone (1980) reported pebble-conglomerates with clasts up to 3 cm in size in the area south of Royal Bay. Craw and Turnbull (1986) described abundant granule and fine-pebble conglomerate on the south side of Ross Glacier. These coarser rocks seem to be identical in composition to the lithic greywackes which form the bulk of the CB.

Table III. Lithofacies scheme used to distinguish Shale (**Sh**), Transitional (**T**), Sandstone (**S**), and Massive Sandstone (**M**) facies, after Tanner and Macdonald (1982)

	Sh	T	S	M
Sandstone/shale ratio	<0.5	0.5–2	>2	>2
Sandstone bed thickness (m)	–	–	<2	>2

The variation of greywacke-shale ratios forms the basis for a lithofacies scheme (Table III) devised by Tanner and Macdonald (1982) in the Fortuna Bay area, and used as a rapid method of "logging" long sections during the course of structural mapping. The three-fold division, into sandstone (**S**), transitional (**T**) and shale (**Sh**) facies, has since been successfully applied all over the CB outcrop (Tanner and Macdonald, 1982; Craw and Turnbull, 1986). Further work on the south-west coast, especially in the area between Larvik and Vestfold Island, led to recognition of the massive sandstone facies (**M**), differentiated from **S** on the basis of bed thickness. Facies occur in units generally 20–100 m thick, but which can be anything from 5–500 m.

Although this facies scheme appears arbitrary it reflects a real division of the CB. Logging within the facies units shows that they can be differentiated in

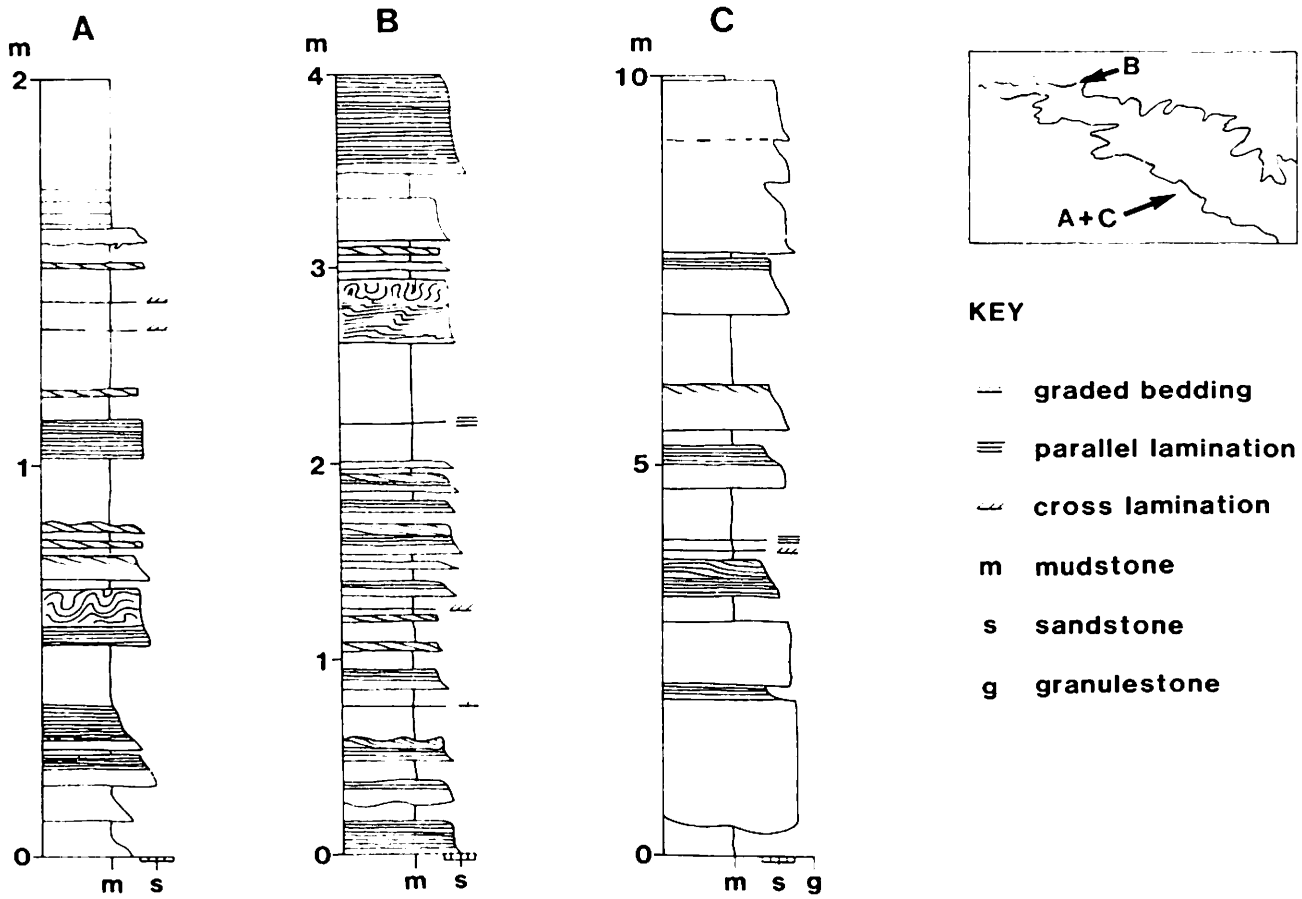


Fig. 35. Typical sedimentary logs showing the major lithofacies of the Cumberland Bay Formation. A, shale facies, Larvik; B, transitional facies, Macdonald Cove; C, sandstone facies, Larvik.

terms of sandstone bed thickness, grain size and sedimentary structures (Fig. 35).

The shale facies (Fig. 36) is typically formed of structureless shale or shale interlaminated with siltstone. Interlaminated with the shale and siltstone are thin, sharp-based, laterally continuous sandstone beds, usually parallel sided and typically 0.5–5 cm thick, rarely exceeding 10 cm. Most sandstone beds are strongly graded, from fine to very fine sandstone, and comprise structureless, flat-laminated and cross-laminated intervals, either separately or in various combinations. Most sandstone beds are clearly turbidites, with only seven types of sandstone bed found out of 243 logged in detail in facies **Sh** (Macdonald, 1982): T_a, T_{ab}, T_{abc}, T_b, T_{bc}, and T_c.

The transitional facies (Fig. 37) consists of turbidite sandstone beds 10–20 cm thick, only rarely exceeding 30 cm, intercalated with shale of the same order of thickness. Sandstone beds usually grade from medium–fine sandstone to fine or very fine sandstone at the top. Lateral continuity is extremely good and most beds have flat bases, though sole-structures of all types are found. Only the same seven bed types as in facies **Sh** were found in 258 beds measured (Macdonald, 1982).

The sandstone facies, **S** *and* **M** (Fig. 37) are divided on the basis of bed thickness, using an arbitrary figure of 2 m; this was purely a mapping convention for splitting thick sandstone sequences. In reality the beds form a continuous spectrum and are described together here. Sandstone beds range from 10–600 cm thick, and are most commonly 20–80 cm. Most beds are parallel sided and laterally persistent at outcrop scale, and some, in the Larvik area, can be traced for 500 m without apparent change in thickness or internal structure. Only 15 examples of lateral thinning in sandstone beds were found by Macdonald (1982); this represents 1.8% of all the sandstone beds measured, and much less than 1% of all beds examined (Macdonald, 1982). Amalgamation is relatively common, most beds have slightly erosive bases and contain shale intraclasts, up to 1 m long, at any level within the bed. Most beds have flat tops and an abrupt junction with the overlying shale.

Macdonald (1982, 1986) recognized a variety of sandstone bed types within the CB. These varied from extremely thick (up to 6 m) structureless very coarse sandstones (Fig. 38), through various poorly graded types to well-graded classical turbidites (Fig. 34). The non-graded and poorly graded beds have few sedimentary structures beyond crude parallel lamination and shale intraclasts (Fig. 39). All of these beds were interpreted as the deposits of high-density turbidity currents, using the criteria of Lowe (1982). Thick ungraded or poorly graded beds are restricted to the south-western coastal outcrop. The classical turbidites are common all over the outcrop, and are perhaps numerically the dominant bed type. Interestingly the T_a division is almost always present and few beds are bottom-absent.

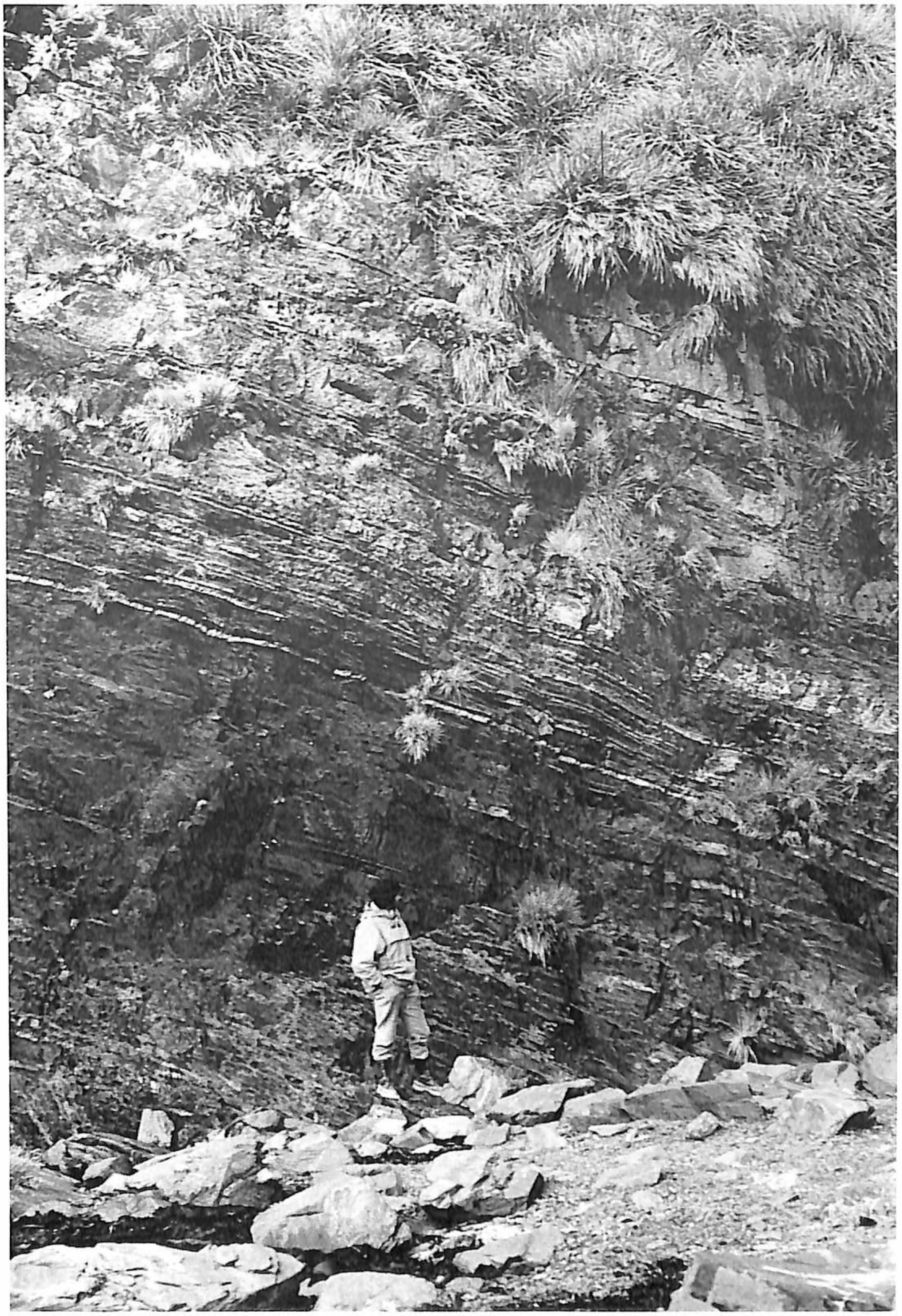


Fig. 36. Typical exposure of the shale facies, Cumberland Bay Formation, coast of Jacobsen Bight.

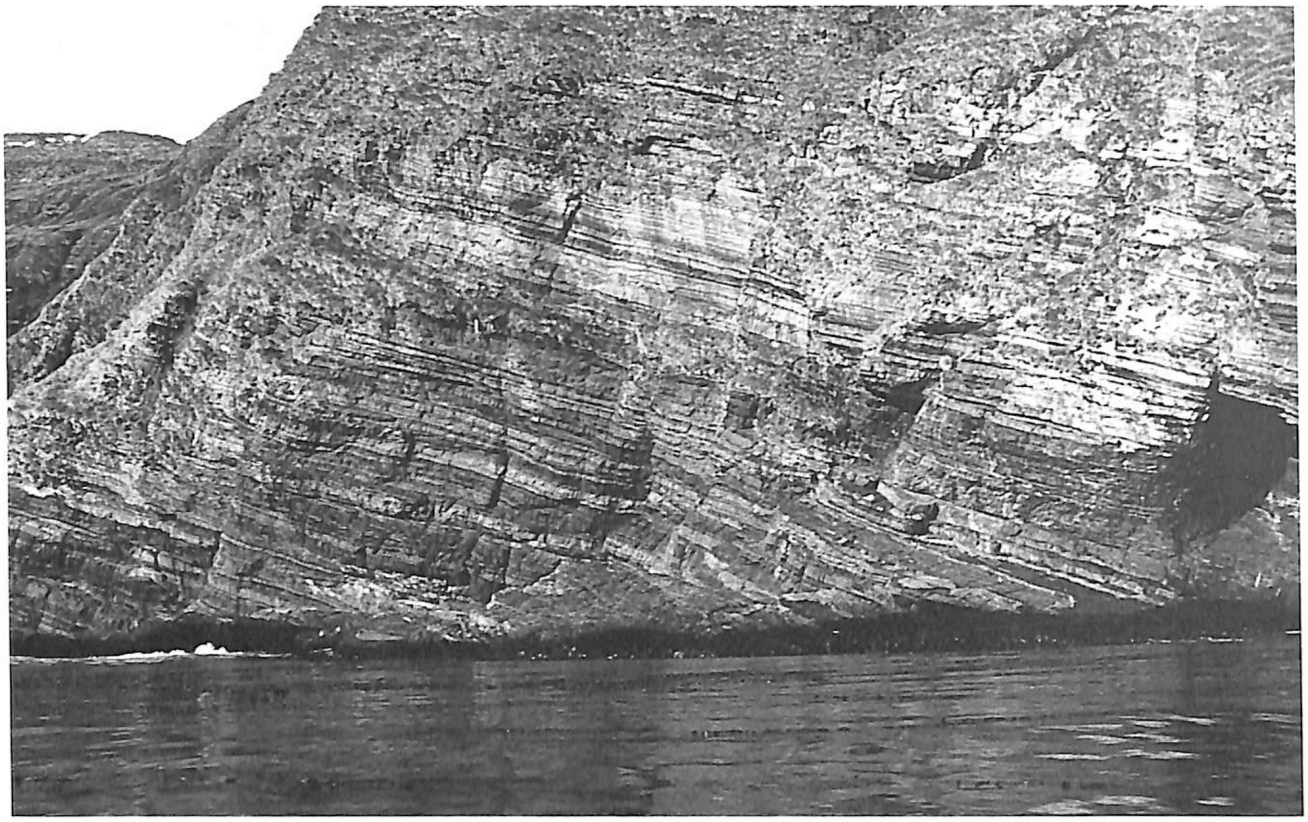


Fig. 37. Transitional facies unit passing upward into thin-bedded sandstone facies, Cumberland Bay Formation, Sappho Point. About 50 m of section are shown.

Facies organization: Facies **S** and **T** predominate in most sections (*c*. 85% of all facies units) which gives most sequences a monotonous appearance (*cf*. Mortimore, 1979). Tanner and Macdonald (1982) recognized a preferred organization of facies into coarsening- and thickening-upward cycles: **Sh** → **T** → **S**. Facies **M** is always completely contained within **S** units, hence the two can be combined for interpretative purposes (Walker, 1984).

Tanner and Macdonald (1982) also recognized a considerable geographical variation in the thickness of facies units and their relative importance in local successions. Along the south-west coast, the facies **S** and **M** form more than 50% of the succession, **Sh** forms less than 20%, and all four facies occur in units 55–60 m thick (Table IV). In contrast, along the north-east coast the massive facies is absent, **S** forms less than 50% and **Sh** is almost always greater than 20% (and may reach 70% of individual sections). Tanner and Macdonald (1982) designated the two areas Zone A (SW) and Zone B (NE), separated by the contour representing 50% of sandstone facies in the succession (Fig. 40D).

The **Sh** and **T** facies are uniform in character across the outcrop of the CB, with no significant change in either mean bed thickness or percentage of various bed types (Macdonald, 1986). In contrast there are strong geographical changes in facies **S** + **M**, summarized in Fig. 40. Mean and maximum bed thickness are higher in Zone A and there is a much higher percentage of amalgamated sandstone beds. Allied with this obvious distinction, there is a more subtle change along the length of the zones, with beds at the north-west end of the outcrop having a much higher percentage of T_b and T_c divisions (Fig. 41). This change is discussed in more detail by Macdonald (1986).

Table IV. Characteristics of lithofacies units in Zone A (south-west coast) and Zone B (north-east coast), after Tanner and Macdonald (1982)

		Sh	T	S	M
Unit mean thickness (m)	Zone A	56	55	58	55
	Zone B	79	43	28	–
Mean percentage of unit within local succession	Zone A	5%	32%	55%	8%
	Zone B	26%	43%	31%	0%

Vertical variation of thickness of facies units, and internal features of the facies such as bed thickness, grain size and sedimentary structures were investigated at each major locality. Although there were obvious limitations imposed on the length of individual sections, a total of 34 sections, ranging in thickness from 111–2163 m, and including 7 sections of more than 1000 m, were measured. The total thickness of strata investigated was 20,400 m with 371 facies units. Beyond the cyclicity mentioned earlier, there was no sustained vertical variation in any section. The character of the facies units is essentially unchanged up any section.

The main sedimentological variation within the CB appears to be geographical rather than vertical. This has

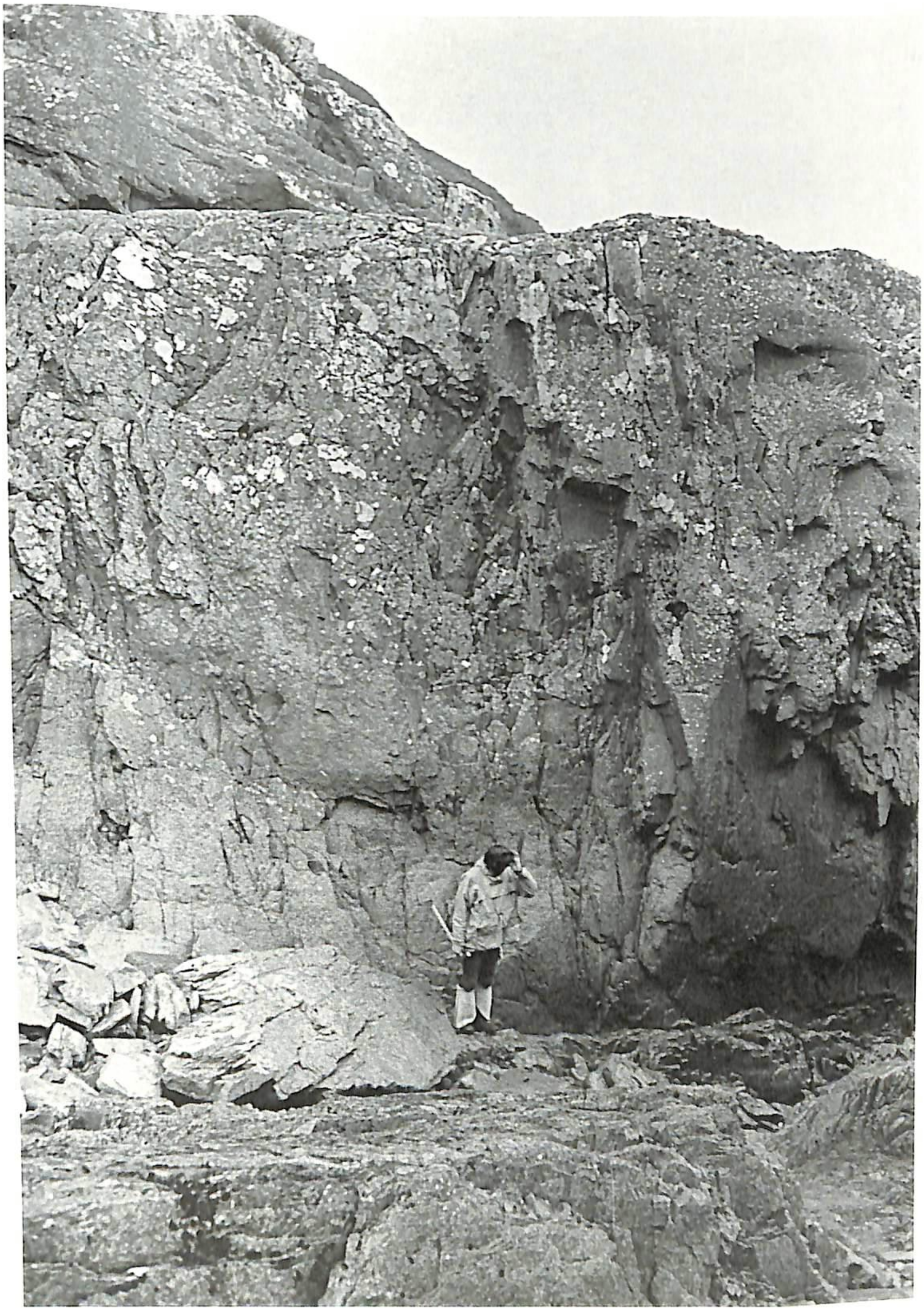


Fig. 38. Single sandstone bed, 6 m thick, of the Cumberland Bay Formation in the Larvik area. This bed is the product of deposition from a high density turbidity current and would be classified as an S_3 bed (Lowe, 1982).

Fig. 39. Elongate shale intraclast in a 1.2 m thick, poorly graded sandstone bed of the Cumberland Bay Formation, Larvik area.

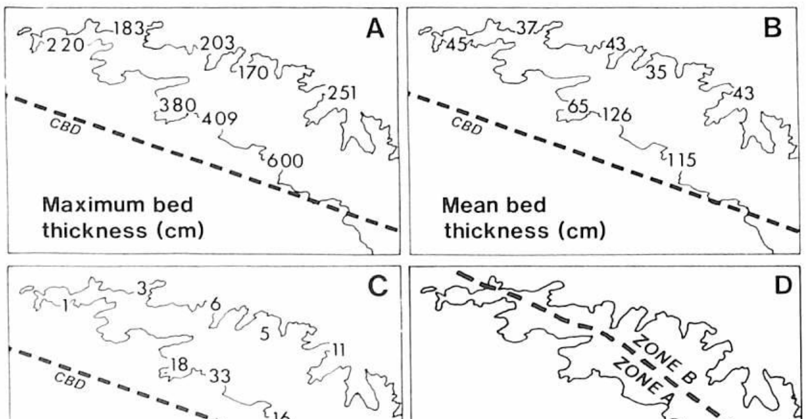


Fig. 40. Variation of (A) maximum sandstone bed thickness, (B) mean sandstone bed thickness, and (C) percentage of amalgamated sandstone beds at eight localities in the Cumberland Bay Formation; the inferred line of the basin margin is shown by the Cooper Bay dislocation (CBD). Map D shows the division of the CB into sandstone-rich (Zone A) and sandstone-poor (Zone B) areas.

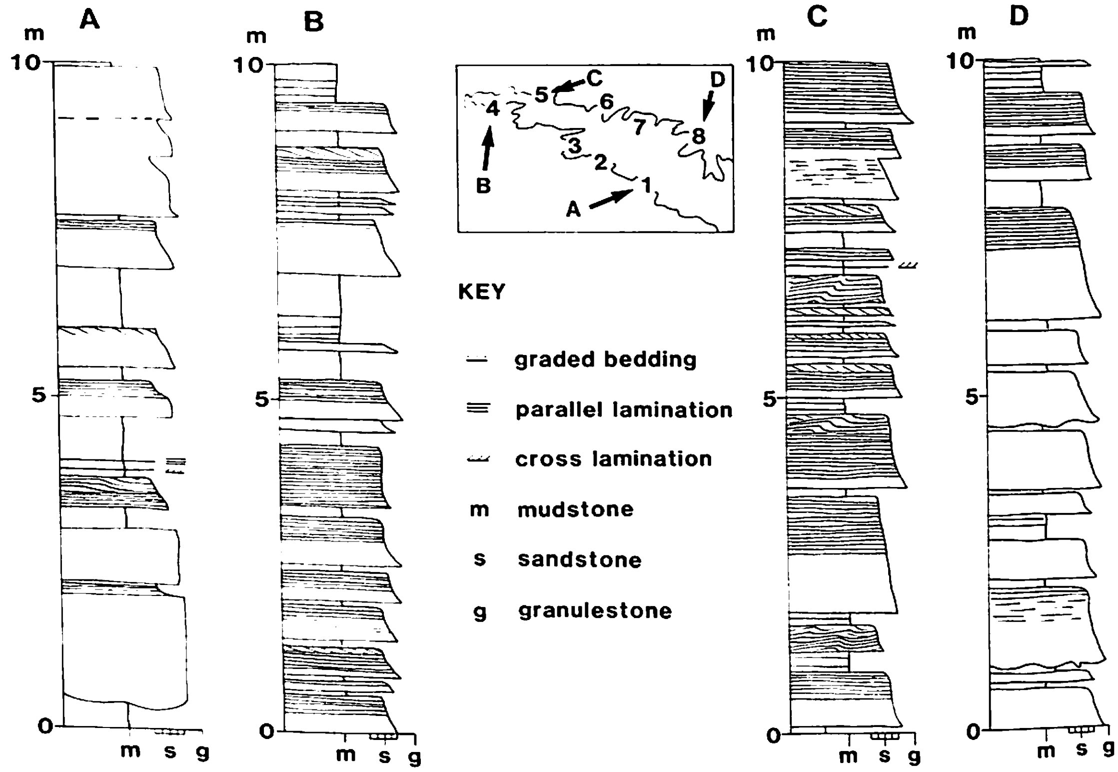


Fig. 41. Sedimentary logs through typical S facies units at four localities in the Cumberland Bay Formation. It is obvious that the beds shown on logs B and C, at the north-western end of the outcrop, have much higher proportions of T_b and T_c divisions than those at localities A and D.

important implications for the study of other turbidite formations, discussed by Macdonald (1986).

Palaeocurrent and palaeoslope analysis: The CB contains a wide variety of palaeocurrent and palaeoslope indicators, discussed in detail by Macdonald and Tanner (1983). Most information comes from sole marks, principally groove casts made by a variety of tools; fossil wood also provides information (Macdonald and Jefferson, 1985). Unambiguous directions come from tools at the end of casts, flutes and frondescent casts. Careful untilting of the field measurements is necessary (Dott, 1974) and failure to recognize the effects of F_2 folding can lead to erroneous directions being calculated (Dalziel and others, 1975; Tanner, 1982*b*; Macdonald and Tanner, 1983).

At every locality in the CB, the principal palaeocurrent direction is towards the west-north-west (Fig. 42). Along the north-eastern coast (Zone B) palaeocurrent distributions are unimodal, but in Zone A every locality has a bimodal distribution, with both west-north-west and north-north-east modes. Macdonald and Tanner (1983) suggested that these represented two source areas: a main axial dispersal direction (directed to the west-north-west), with local lateral supply towards the north-north-east in Zone A. Recent work in the area south of Ross Pass (Craw and Turnbull, 1986) supports these conclusions.

Slump folds are common within the CB, and are locally numerous. Most have amplitude and wavelength of the order of 1–2 m although a few larger examples occur. By using the vergence directions of the folds (*cf.* Woodcock, 1979), Macdonald and Tanner (1983) defined a broadly northerly palaeoslope for the CB. Again, the situation was more complex in Zone A where the slumps defined a palaeo-high in the central part of the zone. This feature was probably lobate and backed against the basin margin (Fig. 42).

Composition: The arenites and minor fine rudites of the CB are lithic volcaniclastic greywackes (in the sense of Pettijohn and others, 1972). Clasts are sub-angular to sub-rounded, with poor to moderate sorting, set in a greenish or brownish grey matrix of fine-grained recrystallized material. There are a large variety of clast types, all of various sorts of igneous rock and fragments of feldspar crystals (Trendall, 1953, 1959; Dalziel and others, 1975; Stone, 1975, 1980; Winn, 1978; Tanner and others, 1981). The lithic:feldspar ratio is strongly dependent on grain size (Macdonald, 1980); the plagioclase: total feldspar ratio is never less than 0.93, and is usually *c.* 0.99 (Winn, 1978). Tanner and others (1981) recognized andesite, altered feldspathic glass, amygdalar andesite, quartz-andesite, dacite, felsite, rhyolite, plagioclase, K-feldspar, amphibole, pyroxene and quartz fragments; most samples had at least trace amounts of all these constituents. Silicic volcanic fragments usually form less than 5% of the clast population and most of the quartz grains (1–5% of clasts) appear to be phenocrysts from silicic volcanic rocks.

Almost all of the samples of CB greywackes plot in the lowest part of a QFL diagram. They are petrographically quite distinct from the Sandebugten Formation greywackes and only the rocks of the Barff Point Member have any appreciable quartz content (Stone, 1982). Other than this there seems to be no variation of petrography with either stratigraphic height or geographical position within the CB.

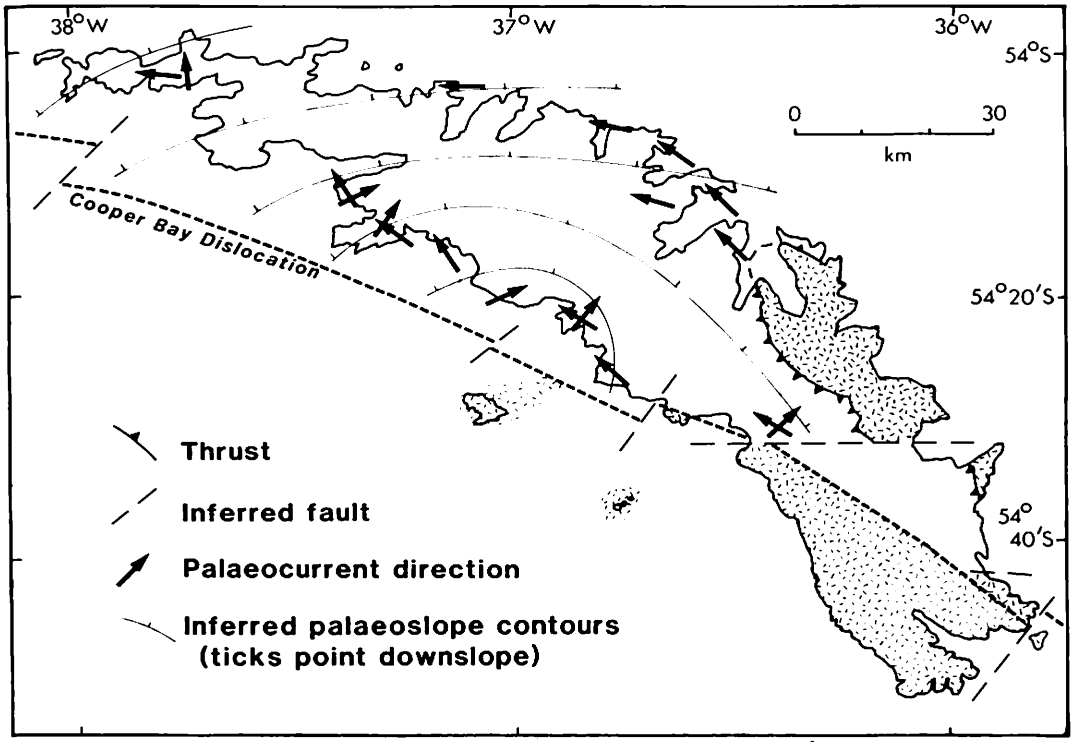


Fig. 42. Summary of the palaeocurrent and palaeoslope analyses from the Cumberland Bay Formation, which is unornamented while all other formations are hachured.

There has been low-grade metamorphism to prehnite-pumpellyite facies (Tanner and Rex, 1979), which pre-dated the formation of S_1 cleavage (Trendall, 1953). Recrystallization of the matrix has occurred with considerable growth of illite, sericitization of feldspar, and the growth of large irregular patches of prehnite and calcite, most strikingly in the widespread early-diagenetic nodules but also in the matrix of most greywackes. Pumpellyite is only present as tiny dispersed specks and is not found in every sample.

Limited geochemistry (Macdonald, 1980; Clayton, 1982) shows that the CB greywackes are markedly calc-alkaline in composition. All the petrographic evidence suggests derivation from a volcanic-arc terrane. There is also a striking similarity with the Yahgan Formation of southern South America (Winn, 1978). These two facts provided one of the lines of evidence which led Dalziel and others (1975) to suggest deposition of the CB/Yahgan formations in a back-arc basin. Under this model the source area of the CB would be represented by the Annenkov Island Formation. Indeed there are close petrographic matches between clasts in the CB and igneous rocks intruding the AF (Tanner and others, 1981). However, Storey and Tanner (1982) have pointed to differences between the geochemistry of the two formations and cautioned against making too close a comparison. The AF probably represents the type of terrane from which the CB was derived, rather than the exact area. This is reinforced by the palaeocurrent evidence which indicates derivation of much of the CB from an area south-east of present-day South Georgia.

Depositional model: The majority of the sandstones of the CB are clearly turbidites, either due to high-density or low-density flows. There is evidence for small-scale reworking by standing currents and minor input by direct volcanic airfall (Storey and Macdonald, 1984), but mass-flows obviously predominated. When compared with other turbidite formations and standard turbidite facies schemes, it is clear that the CB has a very restricted range of facies. Almost all CB rocks belong to facies C, D and B_2 of Walker and Mutti (1973), with no sign of either the finest or coarsest facies. In particular conglomerates and large-scale channels are conspicuously absent and the beds in the CB (even the very coarsest) are laterally persistent. The CB cannot be described in terms of any current submarine fan model, having no channels, a restricted range of facies, no overall progradational trend and a longitudinal/rectilinear palaeocurrent pattern. Although the coarsening- and thickening-upward cycles in the CB are superficially similar to cycles produced by progradation of outer or mid-fan lobes (Mutti and Ricci Lucchi, 1972; Walker, 1978), they are generally thicker and formed of coarser sediment.

Various features of the depositional system have been discussed by Tanner and Macdonald (1982), Macdonald and Tanner (1983), Storey and Macdonald (1984) and Macdonald (1986). Most of the formation seems to have been deposited in a series of unchannelled lobes prograding over interlobe or basin plain areas. These coalesced to form a broad sediment wedge backed against the faulted south-western margin of the basin. High density turbidity currents were locally derived and largely restricted to Zone A, explaining the lateral palaeocurrent directions, thicker coarser beds and complex palaeoslopes. Low density, longitudinal flows operated throughout the basin, flowing towards the west-north-west and accounting for the downcurrent changes in bed character shown in Figs. 40 and 41.

Since the marginal lobes apparently did not inhibit longitudinal flow, deposition must have been matched by subsidence and topography would have been transient, allowing interbedding of axial and lateral deposits. Storey and Macdonald (1984) suggested that due to rapidly shifting sites of volcanism and input into the basin, a true fan system could not develop.

Structural geology

The large-scale folding of the CB is one of the most striking features of the island, used as a navigational aid by mariners and commented upon by early explorers (Shackleton, 1919). Three phases of folding have been recognized over the whole outcrop (Tanner, 1982*b*; Tanner and Macdonald, 1982), but by far the most important of these is the first.

The first (main) phase of deformation involved production of large-scale folds with amplitude and wavelength on the scale of hundreds of metres or more. The axes of these F_1 folds run parallel to the length of the island, as do all the main fabric elements. Along the south-west coast folds are large, upright to slightly overturned, open structures with broad rounded hinge zones and undulating long limbs (the south limbs of the anticlines). In this area folds tend to occur in anticline–syncline pairs with a common limb dipping steeply north-east. The associated axial planar cleavage is rather hackly and anastomosing and does not seem to be significantly fanned, although both divergent and convergent fanning is found in the area around Queen Maud Bay (Macdonald, 1982). Folds of this style pass north-eastwards into continuous trains of close–tight chevron folds with narrow hinge zones and shorter, more planar right-way-up limbs. The common limb of anticline–syncline pairs is vertical or inverted (Fig. 43). Cleavage is much better developed, forming a true slaty cleavage in the shales, with a strong elongation lineation at high angles to fold hinges (Tanner and Macdonald, 1982). In the extreme north-east, divergent cleavage fans are common. The progressive change in style of F_1 is well illustrated by the cross-sections in the inset of the map. The line of onset of chevron folding is quite distinct and can be clearly seen in several mountain ridges in the interior of the island. The change is best seen at Zigzag Pass, on the south side of Esmark Glacier. The S_1 cleavage mirrors this change, decreasing in dip towards the north-east. The L_1 intersection lineation clearly shows that fold axes swing parallel to the island from north-east–south-west in the south to nearly east–west in the Elsehul area. The dominant plunge is west-north-west, although there are local reversals, especially along the south-west coast.

The second phase of deformation is co-axial with F_1 and is restricted to the north-eastern coastal areas. In

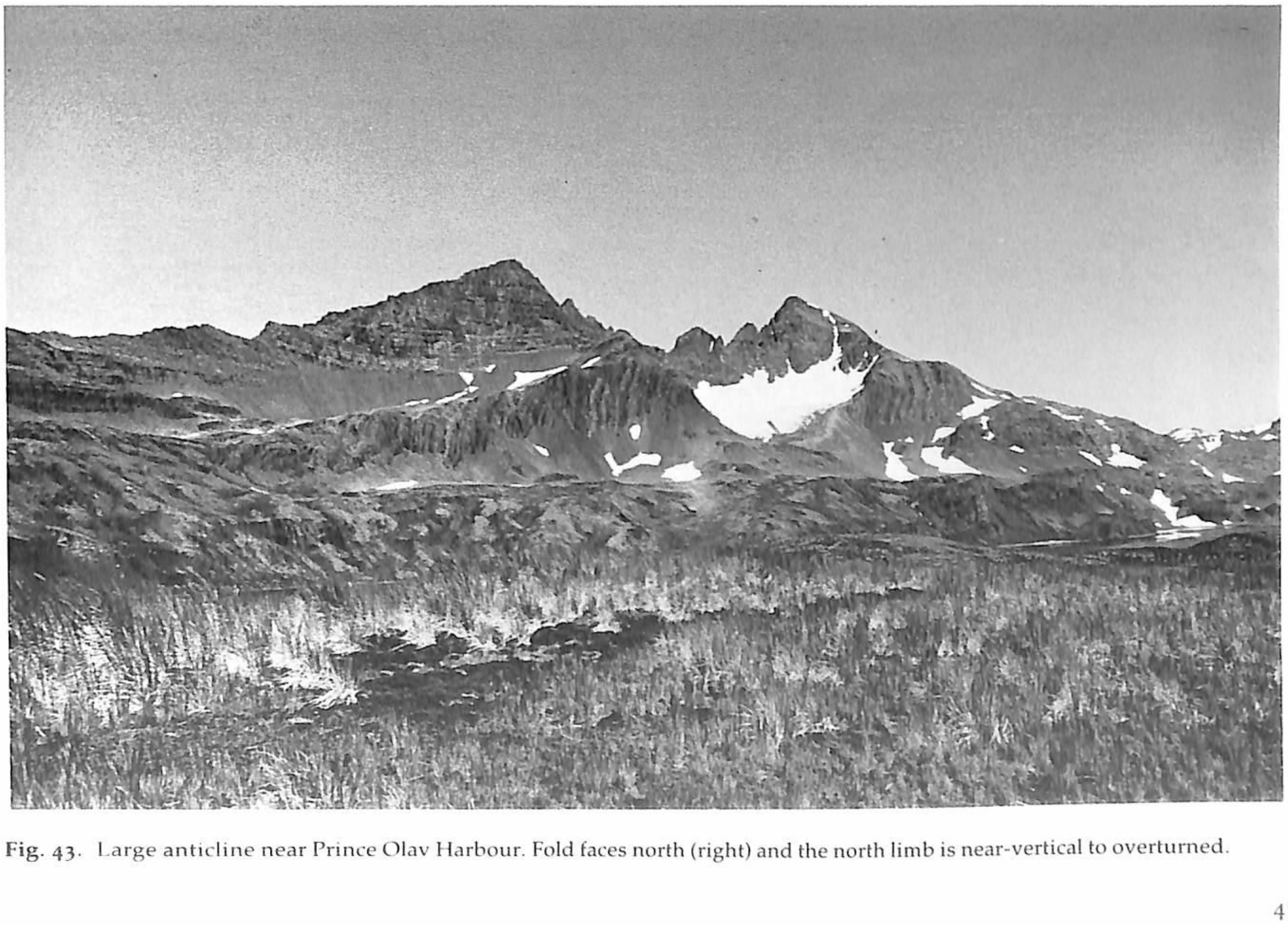


Fig. 43. Large anticline near Prince Olav Harbour. Fold faces north (right) and the north limb is near-vertical to overturned.

Fig. 44. Upright F_2 folds in the Cumberland Bay Formation, Fortuna Bay area. Fold wavelength is about 10 m.

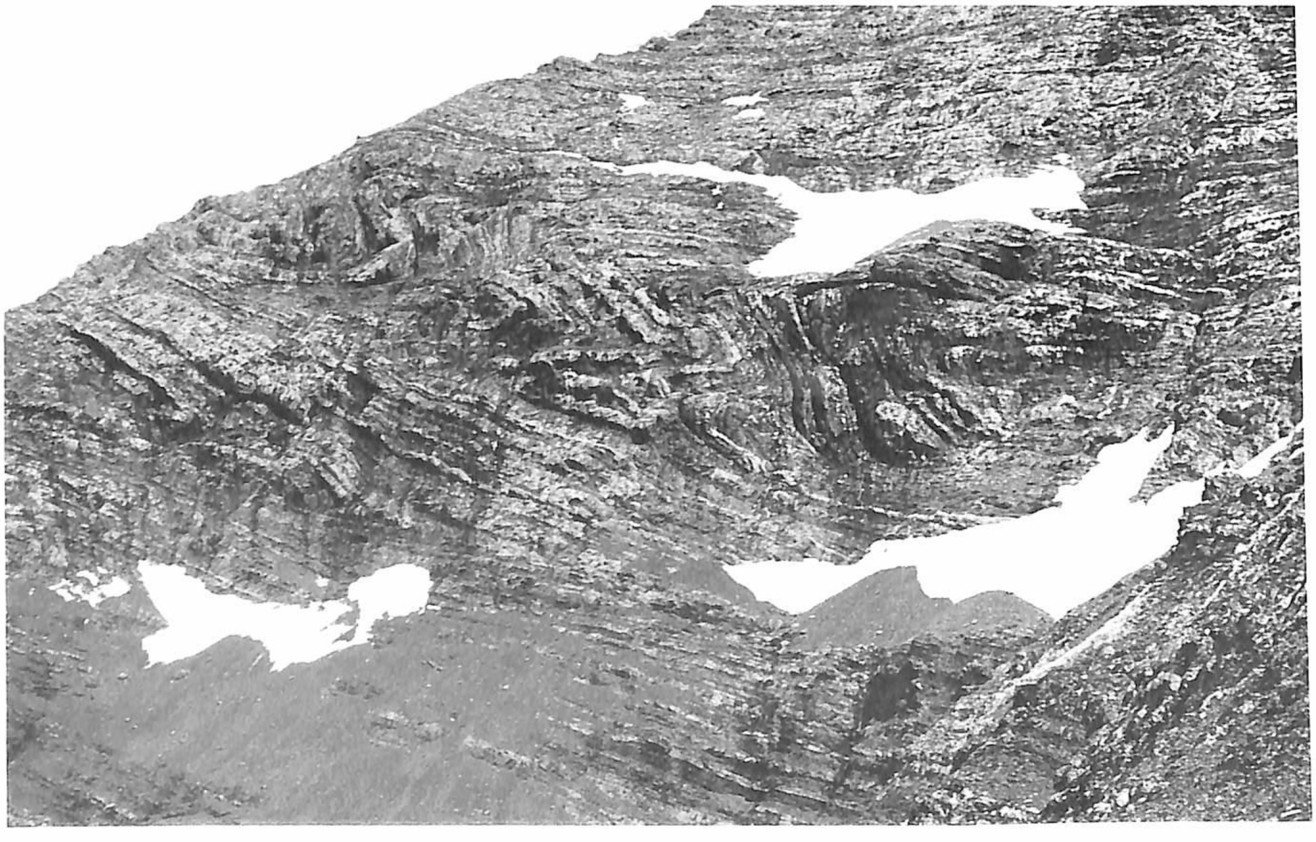


Fig. 45. Large F_1 folds near Fortuna Bay, on the northern edge of a major zone of F_2 folding. Both interlimb angle and axial-plane dip of the first folds have been considerably reduced by the later deformation. About 100 m of cliff are shown.

the field it is apparent as upright open folds, with wavelengths of metres to tens of metres (Fig. 44). In many areas there is a conspicuous vertical crenulation cleavage associated with these folds but both folds and cleavage can be developed separately. In some areas the presence of much larger F_2 structures, with wavelengths of up to 2 km, has been demonstrated by mapping; these have a marked effect on F_1 structures (Fig. 45). In such areas (notably Fortuna Bay, Stromness and near Busen Point) it is impossible to untilt palaeocurrent indicators (Tanner, 1982*b*).

The third phase of folding has no effect on earlier structures, and again is restricted to north-eastern coastal areas. It usually occurs as centimetre-scale kink bands, although a few steeply plunging box folds have been found. Most F_3 structures are oriented roughly north–south.

Other phases of folding have been detected locally. Both Stone (1982) and Macdonald (1982) thought they could recognize a period of buckling pre-dating F_1. Stone (1982) described metre-scale tight folds from the area south of Royal Bay, but evidence that they preceded F_1 is equivocal and they could be a different manifestation of F_2. Macdonald (1982) found open folds with wavelengths of tens of metres apparently transected by S_1. These buckles are on the south limb of a major anticline on the east face of Binnie Peaks and are largely inaccessible. Clayton (1983) suggested that there were very large-scale monoclines parallel to F_3 kink bands, refolding F_1 axes. It is likely, however, that these are an apparent effect due to topography.

Tanner and Macdonald (1982) suggested that the whole deformational history could be due to a simple shear couple acting on the sediment pile. This simple shear was due to south-westerly underthrusting of the basin floor during the mid-Cretaceous.

Igneous intrusions

There are few intrusions into the CB over most of its outcrop, but a number of small pod-like bodies occur in the north-east, all within 1–2 km of the thrust separating the CB from the Sandebugten Formation. The most northerly occurrences, Osmic Hill and Dartmouth Point, have been described by Mair (1981). Here small bodies up to 700 m long by less than 200 m wide are arranged in a discontinuous band, apparently cross-cutting bedding, but parallel to S_1. The igneous bodies have deformed margins and consist of quartz-diorite (porphyry of Tyrrell, 1915, 1916) and gabbro. Stone (1982) reported a small lens of igneous rock (4 m × 50 cm) from Will Point and a much larger occurrence inland from Iris Bay, where there are three sheets of basic igneous rock ranging from 10 cm – 10 m thick exposed over a distance of 3 km. The sheets appear to be conformable with bedding and there have been thermal metamorphic effects on the surrounding sediments. Stone referred to these rocks collectively as epidiorites and showed that they had the original mineralogy of oversaturated dolerites. He pointed out that there were many cobbles and boulders of similar rock in the moraines on Nordenskjöld, Cook and Heaney glaciers and suggested that more sheets cropped out farther inland.

7 Sandebugten Formation

Introduction

The main exposures of the Sandebugten Formation (SF) are south of Barff Point and between St. Andrews Bay and Royal Bay. There are also large exposures at Cape Charlotte and Dartmouth Point, with smaller areas at Will Point and on the northern and eastern ridges of Mount Brooker. In general the areas underlain by the SF are lower and less rugged than the topography of any of the other major units, with no peaks above 1000 m.

Trendall (1953, 1959), Aitkenhead and Nelson (1962), Dalziel and others (1975), Stone (1975, 1980) and Tanner (1982*b*) have all worked on the Sandebugten Formation, but of these, only Stone worked in any detail on the formation.

Stratigraphic and age relations

Neither base nor top of the SF is known and the whole unit lies below a major thrust plane which separates it from the CB (Fig. 46); there are no boundaries between the SF and any other unit. The SF is correlated with the Cooper Bay Formation and the Inland Member of the Ducloz Head Formation on petrographic grounds (Storey, 1983*b*).

No fossils except small fragments of carbonized wood have ever been recovered from the SF. Burrows of various sorts, principally *Chondrites*, have been found but these have no age significance.

Various attempts have been made to date the SF on petrographic grounds. In particular it has been compared with the supposed Upper Palaeozoic "Trinity Peninsula Series" (now Trinity Peninsula Group) of the Antarctic Peninsula (Trendall, 1953; Adie, 1964; Skidmore, 1972). Subsequently Trendall (1959) suggested that the SF was more probably Mesozoic but was older than and deformed before the Cumberland Bay Formation. Dalziel and Dott (1973) and Dalziel and others (1975) thought that the CB and SF were facies variants deposited at the same time on opposite sides of a back-arc basin. Stone (1980) supported this view but thought that the main phase of deformation was earlier in the SF than in the CB. Radiometric dating of six samples of slates from the Sandebugten Formation yielded ages of 66 ± 3 – 84 ± 3 (Thomson and others, 1982) (recalculated to 68–86 Ma using current conventional constants). These were interpreted as uplift and cooling ages and are well within the range derived from the CB. Thus at least the latest part of the structural evolution of the two formations was coeval.

Stone (1980) defined a type section of the SF from a cliff section on the coast 0.5 km north-west of Sandebugten. His log (Stone, 1980, table 1) shows 35 m of strata thought to be typical and should be regarded in the same light as the type sections of the CB defined previously.

Sedimentology

Lithology and facies: The SF consists of a thick sequence of alternating greywacke and shale with minor fine rudite. The coarsest sediment is pebbly granulestone with clasts up to 6 mm. Beds are laterally persistent and most show normal grading, although structureless beds are relatively common and inverse grading has been reported (Stone, 1980). The upper parts of a substantial minority of beds show parallel and cross lamination, and the beds have been interpreted as turbidites (Stone, 1975, 1980). In general the unit is finer grained than the CB, and is certainly thinner bedded, although Stone (1980) refers to beds up to 1.45 m thick.

Trendall (1953) recognized that there were two major lithological associations in the SF: in sections of up to 20 m either coarse, or fine rocks always predominated. Beyond this observation there has been no further work on facies of the SF. Craw and Turnbull (1986) reported the presence of significant chert and marble bands interbedded with quartzo-feldspathic schist on the east side of Heaney Glacier. These lithologies were previously unknown in the SF.

Palaeocurrent analysis: Most of the limited sedimentological work has concentrated on the determination of palaeocurrent direction. Trendall (1959) suggested a source to the east-north-east, on the basis of six measurements of cross lamination. Dalziel and others (1975) and Winn (1978) measured 214 cross laminae and found a strongly bimodal pattern with a north-north-east to south-south-west trend. The idea of a north-easterly source gained further support from Stone (1980) who measured four sole marks which had a south-westerly mean. This was in marked contrast to six sole marks measured by Tanner (1982*b*), with a west-north-west mean, well within the field of readings from the CB.

It is obvious that the palaeocurrent pattern in the SF is different to that in the CB. However, the biggest single set of readings, those obtained by Dalziel and others (1975), should be treated with extreme caution for a number of reasons:

(a) The precisely bipolar, bimodal distribution is extremely unlikely, if not impossible to achieve in a deep-marine turbidite formation.

Fig. 46. Thrust contact between steeply dipping Sandebugten Formation (below) and more gently dipping Cumberland Bay Formation (above) at Will Point. Sea stack is about 10 m high.

(b) Each mode is extremely tight, much more so than that obtained in the CB by the same authors. By its very nature cross lamination tends to give highly dispersed readings (*cf.* Bluck, 1974).

(c) The mean trend through the two modes is precisely perpendicular to the structural trends in the area of Barff Peninsula where Dalziel and others (1975) worked. This evidence tends to suggest a considerable degree of tectonic re-alignment of the cross laminae.

Composition: The SF arenites are all lithic greywackes (Pettijohn and others, 1972). The clast population is radically different from that in the CB, and comprises quartz, polycrystalline quartz, igneous rock fragments, and feldspar crystals. The clasts of igneous rock are principally felsites with subordinate trachyte and granitic clasts (Stone, 1980).

The SF is distinct from the CB, not only in its higher percentage of quartz, but also in the different lithic grains. The provenance of the unit was probably a continental area with silicic volcanic rock; Dalziel and others (1975) and especially Winn (1978) drew attention to the Jurassic Tobífera Formation of southern South America as the probable source. The petrographic evidence for a northerly or north-easterly source is much stronger than any of the palaeocurrent data.

Metamorphism of the Sandebugten Formation was to lower greenschist grade (Tanner and Rex, 1979), with a dominant stilpnomelane-epidote-(prehnite) assemblage (Tanner, 1982*b*).

Depositional model: There are insufficient data to model the SF properly. The deposits are clearly turbidites, which probably accumulated in deep water. The lack of fossils and the sparsity of the reported ichnofauna suggest conditions hostile to any bottom fauna.

The petrographic distinction between the SF and CB is striking. Despite reports of minor amounts of CB-type greywacke within the SF and *vice versa* (Winn, 1978), it is clear that there is little mixing of the two lithologies. There are three main possibilities to account for this:

(a) The SF represents an earlier, and quite distinct unit from the CB, with a continental provenance.

(b) The SF is an early facies of the CB, representing the deposits of an early phase of continental rifting during development of the basin.

(c) The SF and CB are true time-equivalents with separate provenances.

Most authors favour (b) or (c) or a combination of both. If the two formations are time equivalents there must have been some mechanism to prevent mixing of the arc-derived CB and the continentally derived SF. This could either be due to a large distance between the two sides of the basin or to a physical barrier. Macdonald and Tanner (1983) suggested either a raised block of continental crust in the basin floor or an axial rise; they favoured a transient feature such as a rise on tectonic grounds.

Structural geology

The whole of the Sandebugten Formation is much more strongly deformed than the CB. Stone (1980) recognized three major phases of folding. F_1 produced a series of tight chevron folds with wavelengths of tens of metres, and a penetrative fabric which is a schistosity in places (Craw and Turnbull, 1986). On Dartmouth Point these folds plunge gently north-west. Along a north-easterly traverse through this area there is a change in facing direction, from south-west on Dartmouth Point to north-east at Reindeer Valley. This transition appears to be gradual, and the overall structure resembles an anticlinorium (Stone, 1980; Tanner, 1982*b*). These F_1 folds are an order of magnitude smaller than the F_1 folds of the CB, and the change of facing direction could be due to the effects of later folding.

The second phase of folding increases in intensity southwards, and is largely absent in the northern part of the outcrop, near Barff Point (Stone, 1980). The most obvious effect of D_2 is a sub-horizontal crenulation cleavage, which refolds S_1. There are associated minor folds with east–west-trending hinges and an axial-plane separation of up to 3 m. The F_2 structures are best developed in the vicinity of Whale Valley, on the north side of Royal Bay, where large recumbent folds with wavelengths of tens to hundreds of metres can be seen to refold F_1 axial planes (Stone, 1980, fig. 13, plate III).

Development of the third phase of deformation is patchy, but is much stronger in the most southerly outcrops of the SF. It is present as a very fine sub-vertical, north-north-east-trending crenulation cleavage (S_3). No F_3 folds have been seen, but Stone (1980) inferred their presence from major re-orientation of S_2 in the Cape Charlotte area. He envisaged very large-scale open folds trending north-north-east.

The timing of structural events in the SF relative to the CB is conjectural at present.

Igneous intrusions

In the area just north of Gold Harbour, Stone (1980) recognized three igneous bodies tens or hundreds of metres long and metres to tens of metres thick. These rocks are conformable with the local planar fabric, and form large boudins. They lie just below and parallel to the thrust plane.

Petrologically they are ophitic dolerites and epidiorites, identical to the igneous rocks from the Cumberland Bay Formation.

8 Cooper Bay Formation

Introduction

The Cooper Bay Formation (CO) is a sequence of highly deformed volcaniclastic metasedimentary rocks and metabasites that crops out on the eastern side of the Cooper Bay dislocation zone (CBDZ). It forms the south-eastern promontory of South Georgia between Cape Vahsel and Cooper Bay, and extends as far north as Twitcher Glacier where a faulted or thrust boundary with the Cumberland Bay Formation is inferred. The Cooper Bay Formation separates the CB from the DFC.

The sedimentary and meta-igneous rocks were origi-

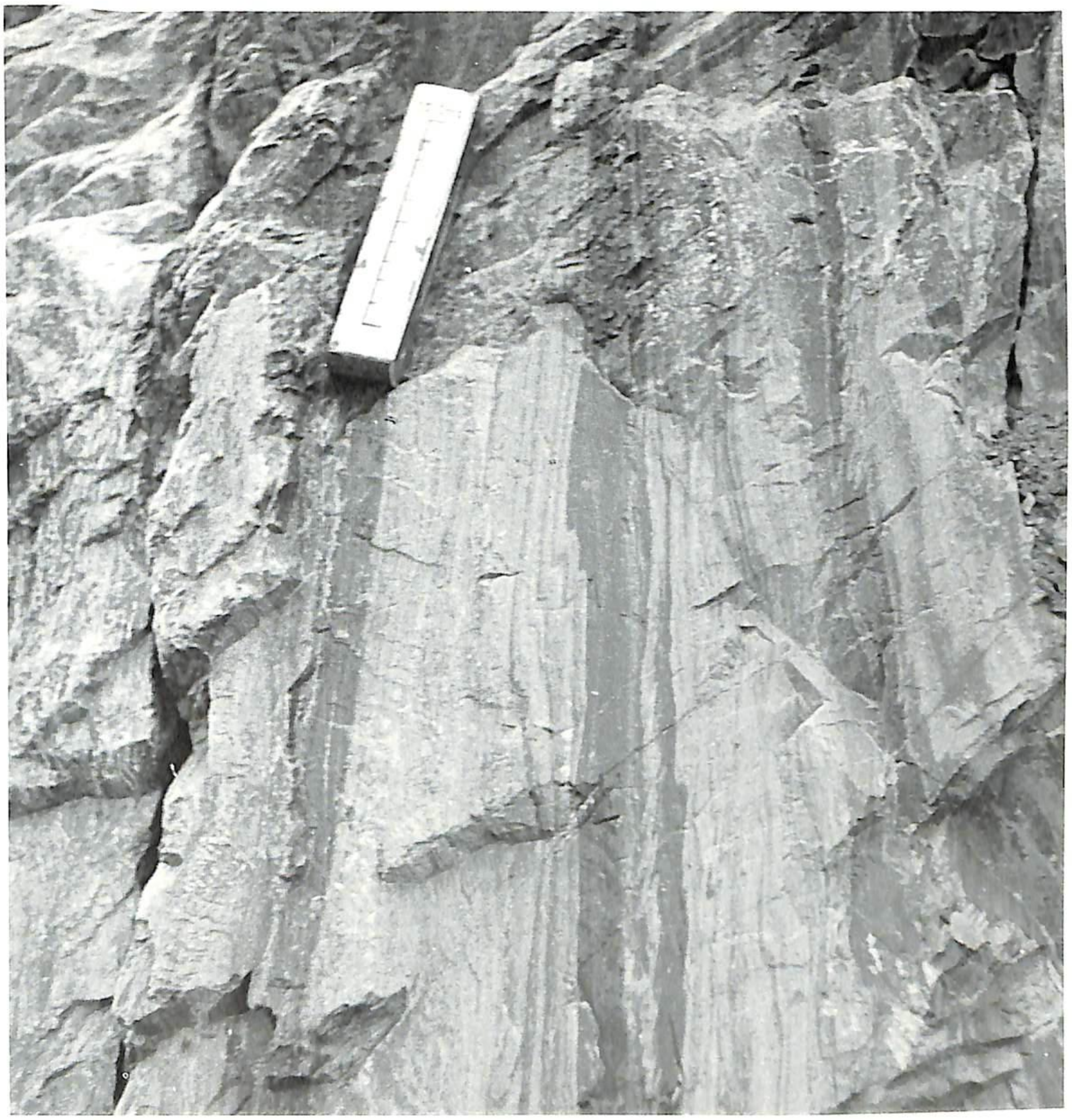


Fig. 47. Mylonites within the Cooper Bay Formation. The scale is 10 cm long.

nally described by Stone (1982) as the Cooper Bay metasediments and were formally termed the Cooper Bay Formation by Storey (1983*a*). The sedimentary rocks are derived from a continental volcanic-arc terrane and are similar to the Sandebugten Formation. Although there is no control on the age of these rocks they are considered to be facies variants of the CB and SF (Stone, 1982) and to be part of the infilling of the Mesozoic back-arc basin.

Metasedimentary rocks

The metasedimentary rocks are an intensely deformed, interlaminated and interbanded sequence of grey, green and brown metagreywacke, schist, phyllite, slate and mylonite (Storey, 1983*a*). The banding is sub-parallel to a penetrative biotite-chlorite-muscovite schistosity and to thin, deformed quartz-rich segregation lamellae or boudins. Sedimentary structures have mostly been destroyed but graded sandstone beds up to 1 m thick occur to the south of Twitcher Glacier. Recrystallization has destroyed much of the clastic constituents of many of these metasediments and only isolated quartz and feldspar crystals and few lithic fragments are recognizable. The lithic clasts are derived from felsic volcanic rocks with some intermediate lithologies. Within discrete zones of high deformation the metasedimentary rocks are finely banded green and white mylonites (Fig. 47).

Metabasites

The metabasites (Storey, 1983*a*) are laminated dark green schists and phyllites and boudinaged green coarse-grained sheets up to 9 m wide within the metasedimentary rocks. They are often difficult to distinguish from the metasediments and may have been emplaced as mafic sills within the sediments prior to deformation. They vary in composition from gabbro to diorite and show a well-developed ophitic texture.

Fig. 48. Finely banded folded mylonites within the Cooper Bay Formation. The scale is 10 cm long.

Structural history

In the metasedimentary rocks of the CO, four distinct fold episodes (Fig. 48) have been recognized (Stone, 1982; Storey, 1983*a*). The first three were, for the most part, co-axial, trending north-west and may represent progressive strain within a deformation event.

The relationship of the deformation of the mylonites in the CBDZ is not certain. In the CO, mylonites are associated with zones of high strain during F_2 folding, and earlier mylonitic fabrics have also been recognized (Stone, 1982; Storey, 1983*a*). Stone considered that myloblastites formed in response to local cataclasis between F_1 and F_2, and Storey considered the finely laminated fabric folded by F_1 and F_2 interference fold patterns to be mylonitic. It is thus likely that the F_1–F_2 fold episodes of the CO are associated with ductile shearing and that complex folding and refolding of contemporaneous mylonitic fabrics occurred within the same shearing event (Fig. 48). A fabric analysis by Dr A.W. Meneilly during a short visit to the area in 1985, suggested a major strike-slip component of shearing in the CO.

9 Ducloz Head Formation

Introduction

Ducloz Head forms the northern promontory of Undine South Harbour, on the south-west coast of South Georgia. The area was visited during a six-day period in the austral field season of 1975–76 (Storey, 1983*b*). Ducloz Head is unique within the context of South Georgia geology; it occurs at the junction between the CB, the LHC and the AF, and has its own characteristic range of lithologies, some of which can be correlated with part of the AF, the LH_1 and the SF whereas others are not found elsewhere on South Georgia. The sedimentary rocks of the Ducloz Head Formation (DH) are divided into two subdivisions: the Coastal (DH_2) and Inland (DH_1) members, the former of which contains small mafic plutons with sheared margins. The members are separated by an inferred fault and the Inland Member is separated from the andesitic greywackes of the CB by a fault zone, possibly a continuation of the CBDZ.

Sedimentary and volcanic rocks

Coastal Member: The coastal member consists of massive sub-quartzose epiclastic sandstone and sandstone breccia up to 10 m thick interbedded with thin-bedded sandstone, siltstone and black shale, basaltic pillow lava and white and buff-coloured massive volcaniclastic breccia, thin-bedded graded tuff and interlaminated tuffaceous mudstone (Storey, 1983*b*). The coarse-grained sandstone and breccia beds are structureless or crudely graded with angular to sub-angular clasts, up to 6 cm in diameter, and very poor sorting. The finer-grained units occur as disorientated sheared blocks between the massive sandstone and breccia and commonly exhibit graded bedding, bioturbation, ripple marks, fine cross laminations, and load and flame structures.

The epiclastic sandstones are quartzo-feldspathic arenites and greywackes comprising up to 80% rounded to sub-angular quartz and feldspar crystal fragments and up to 10% lithic clasts (granitic fragments, porphyritic felsites, muscovite-chlorite-schists and quartz-feldspar-gneiss). The sandstone breccias contain angular sandstone fragments up to 10 cm in length in an unsorted sandy matrix and are believed to be of intraformational origin. The volcaniclastic felsitic breccias contain angular to sub-rounded white and pink-coloured quartz- and plagioclase-phyric rhyolite and dacite clasts and quartz and feldspar crystal fragments. The crystal-lithic tuffs and tuffaceous mudstones have a similar mineralogy and often contain brown circular and ellipsoidal (?) spores which may have a reticulate surface pattern.

The massive unsorted character of the thick breccia and sandstone units suggest that these sedimentary rocks may have been deposited by debris flows in a rapidly subsiding basin. It may have been fault bounded, contained some mafic volcanism and was close to a felsic volcanic source.

Inland Member: The Inland Member consists of highly deformed, poorly preserved interbedded yellow to dark brown graded tuffs, 3–12 cm in thickness, and thin-bedded cherty-looking black radiolarian-rich mudstones. Cross lamination is present in some of the thin-bedded horizons and carbonized plant remains and numerous calcareous nodules are found within the tuffaceous horizons.

The tuffs are crystal-vitric and lithic varieties with angular plagioclase and subsidiary quartz fragments, vitric shards and microlitic feldspar clasts in a devitrified often zeolitized (analcite) glassy matrix.

The mudstones represent normal hemipelagic subaqueous sedimentation and the graded-tuff units formed during periodic introductions of pyroclastic material from an intermittent volcanic source. The Inland Member is similar to, and has been correlated with, the Lower Tuff Member of the AF (Storey, 1983*b*).

Mafic intrusions

Fine- to medium-grained green mafic igneous rocks occur as sheared blocks up to 30 m wide within the deformed sedimentary rocks of the Coastal Member and a single mafic dyke intrudes them. The mafic rocks are similar to the basaltic pillow lavas and are probably co-genetic. They contain primary ophitic augite in a chlorite-albite-epidote-prehnite-calcite-rich groundmass. Sphene and opaque phases are disseminated throughout.

Structural history

The Coastal and Inland members of the DH are both highly deformed, the chaotic nature of which led Storey (1983*b*) to describe the area as a broken formation. Within the Coastal Member, sheared blocks of folded and boudinaged sandstone and shale units occur in disorientated massive sandstone breccia and mafic igneous units. The large-scale folds, which are mainly close with angular hinge zones, are disrupted by shear-

Fig. 49. Disrupted fold of the interbedded tuffs and mudstones of the Inland Member, Ducloz Head Formation. The scale is 10 cm long.

ing (Fig. 49). Fold hinges are variable in trend and plunge, possibly due to dislocation and disorientation of the folded blocks, and there is no well-developed fabric in these rocks. A similar style of deformation has affected the Inland Member: large-scale asymmetric close folds with angular and sub-rounded hinge zones are present; fold-hinge orientations, which plunge towards the north-west are more consistent than in the Coastal Member. Fragmented and disorientated tuffaceous laminae occur within some hinge zones.

10 Cooper Bay dislocation zone

Introduction

The Cooper Bay dislocation zone (CBDZ) is a major tectonic and lithological break [Cooper Bay mylonite zone of Bell and others (1977)] that separates the Drygalski Fjord Complex (DFC), part of the basin floor, from the sedimentary fill. It represents a zone of high strain, extending from Cooper Bay in the south to Spenceley Glacier in the north and consists of a narrow belt, less than 1 km wide, of mylonitized sedimentary, igneous and metamorphic rocks of the DFC. The fault zone separating the Inland Member of the Ducloz Head Formation (DH) from the Cumberland Bay Formation (CB) may represent a continuation of this zone which was displaced in a sinistral sense along an inferred fault through the Ross Glacier–Ross Pass–Brøgger Glacier area. It is also possible that the deformation of the DH described as a broken formation (Storey, 1983*b*) occurred within this dislocation zone, and that the whole of the formation lies within the zone. To the south the dislocation zone must also be displaced by a major transverse strike-slip fault as it does not occur, along strike, on Cooper Island. A marked topographic break separates the dislocation zone from the Cooper Bay Formation (CO) and, although mylonites occur on both sides of this line, the predominance of igneous protoliths of the DFC within the zone are used to separate it from the volcaniclastic sedimentary rocks and mafic sills of the CO.

Within the dislocation zone, narrow bands of laminated green and white mylonite and ultramylonite separate areas of weakly mylonitized and undeformed gabbro, metagabbro, mafic dykes, quartz-diorite, granodiorite, porphyritic felsite and folded metasedimentary rocks of the DFC. The early history of these rocks is similar to that of the remainder of the complex; in the southern part, large gabbro bodies cut by mafic dykes intrude folded metasedimentary rocks of the Salomon Glacier Formation. Quartz-diorite, granodiorite and porphyroblastic gneiss intrude, net vein and form migmatitic relationships with the mafic rocks. The migmatization within the CBDZ (in contrast to that within the DFC) is accompanied by growth of large plagioclase and K-feldspar porphyroblasts in the metasedimentary and igneous rocks. Subsequent mylonitization has resulted in a porphyroblastic gneiss, which is a characteristic lithology of the zone. In the northern part of the zone there are outcrops of porphyritic felsite and volcaniclastic felsite of the Novosilski Glacier Formation.

Structural history

The zone has had a complex structural history with a main mylonite fabric superimposed on the premagmatic fold history of the metasedimentary rocks and on the magmatic rocks. This resulted in a sequence of metabasites, mylonitized granitoid gneisses, quartz-feldspar mylonites and ultramylonites; where deformation is intense the original sedimentary or igneous nature of these rocks has been destroyed and it is possible that protoliths of the CO may occur within the zone. As well as the mylonitization, minor conjugate shear zones, conjugate faults and post-mylonitization folding are common within the zone.

The mylonite fabric defined by alternating quartz and epidote-rich bands is a strong north-west-trending planar surface (Fig. 50) sub-parallel to the margins of the zone. It is mainly present within the granitic rock, whereas the gabbro and mafic dykes form sheared and boudinaged lenses within the mylonites and granitoid gneisses. The mylonites contain discontinuous pale green, white and black colour banding, lineations, intrafolial folds and porphyroclasts.

The minor shear zones are zones of intense deformation aligned both normal and parallel to the margins of the dislocation zone, which show both sinistral and dextral displacement. The folding of the mylonite fabric is often concentrated within these zones and also occurs along the margins of the dislocation zone. The earliest folds are intrafolial isoclinal folds which were probably associated with the formation of the mylonite fabric, or may represent sheared out remnants of the pre-mylonite folds. The F_1 folds of the mylonite fabric vary from close and tight similar folds to small-scale crenulation folds which plunge gently to both the south-east and south-west and have steeply inclined axial surfaces; these are probably conjugate pairs formed during a single deformation event. Within zones of intense deformation the F_1 folds have curved hinge lines and may plunge steeply. These are probably sheath folds formed by rotation of the fold hinges during progressive shear. A late-stage open fold (Fig. 50) about a north-east axis may account for the variation in planar and linear data throughout the zone.

The dislocation zone is consequently a zone of high strain which formed along the margin of the DFC. The DFC remained for the most part undeformed and much of the deformation was taken up in this one zone. The fabrics, folds and minor shear probably formed during a major phase of ductile shearing, the kinematics of

Fig. 50. Folded mylonites within the Cooper Bay dislocation zone, near Cooper Bay.

which are not fully understood. The mylonite fabric and both intrafolial and later folds probably developed by progressive shear and rotation of planar elements during shearing. A preliminary analysis of shear criteria within this zone by Dr A.W. Meneilly has shown a major component of the shearing to be a north-easterly direction, on a steeply inclined reverse fault, as suggested by Storey (1983*a*) and Tanner and Macdonald (1982). This is in marked contrast to the sense of shearing within the neighbouring CO where D_2 mylonites were formed during strike-slip movement.

Although the full history of the movement within these zones is not known, the CBDZ probably formed, and the CB was deformed, during the closing of the back-arc basin; the DFC, part of the floor of the basin, was uplifted and thrust towards the continental margin whereas the basin fill (CB) was underthrust beneath the basement block. The transcurrent movement apparent in the CB may be responsible for juxtapositioning the different tectonic elements of the back-arc-basin system.

11 Marine geology

There has been very little geophysical work on the South Georgia block itself, as most of the research effort has been concentrated on the history of development of the Scotia Sea (British Antarctic Survey, 1985). However, the overall form of the microcontinent is reasonably well known. Simpson and Griffiths (1982) presented the results of bathymetric, magnetic and gravitational studies, and a single unreversed seismic refraction line.

The bathymetry shows that the surface of the block is relatively smooth, with major channels kilometres to tens of kilometres wide and over 100 m deep. These submarine valleys correlate with major glaciers onshore, and reach out as far as the shelf edge, which falls off steeply from the 500 m isobath to abyssal depths. The channels are considered to be glacial in origin (Simpson and Griffiths, 1982).

The magnetic anomaly field is divided into two contrasting regimes by a line trending west-north-west across the block. To the north-east of this line the anomalies are low amplitude, long wavelength and principally negative while the south-west part of the block is characterized by short-period, high-amplitude anomalies with strong positive peaks. The largest anomalies are concentrated along the extreme south-western edge of the block. The line separating the two fields can be traced from the west-north-west, trending towards Cape Darnley. South-east of the island it can be picked up just north of Cooper Island and traced to the east-south-east, passing just north of Clerke Rocks. This line correlates exactly with the on-land Cooper Bay dislocation (Simpson and Griffiths, 1982).

The gravity data indicate a generally high Bouguer anomaly (at +1000 gu) and there is a pronounced west-north-west trend to the anomalies. There is a strong high on the south-west edge of the block, separated by an irregular band of lows from another belt of high values which extends from the northern edge of the block, through Annenkov Island and the Pickersgill Islands, towards Cape Disappointment. This trend is interrupted by a large low anomaly south of the island and then reappears as a +2000 gu high south of Clerke Rocks. On the north-eastern shelf, the gravity values are generally low, although the distinction between north-east and south-west is less marked than with the magnetic data.

Simpson and Griffiths (1982) compared a magnetic profile across South Georgia with one across the Andean cordillera south of Punta Arenas. The striking similarity between the two profiles led them to suggest that a large magnetic body, similar to the Patagonian Batholith, underlies the south-western margin of the South Georgia block. This is an important conclusion, which further underlines the close similarities between South Georgia and southern South America. They interpreted the rest of the geological structure with reference to the units known on land (Fig. 51) and in particular

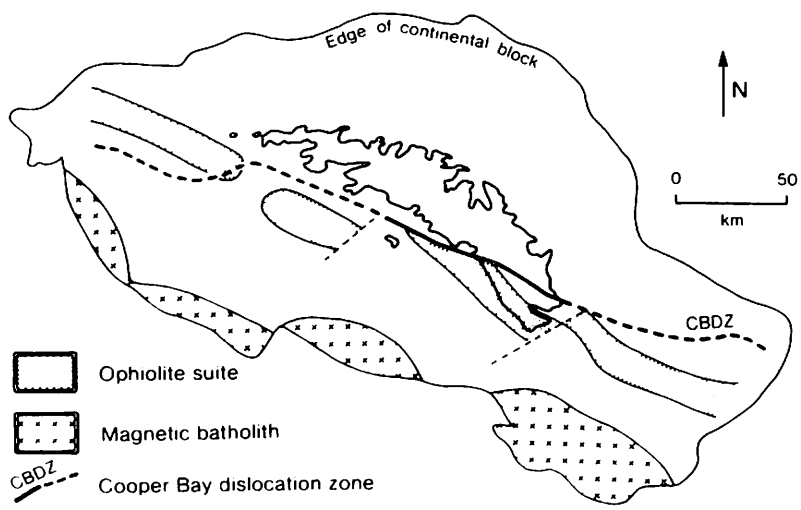


Fig. 51. Interpretation of the major geological features of the South Georgia continental block (after Simpson and Griffiths, 1982).

suggested that a belt of ophiolitic rocks characterized by a relatively weak magnetic signature and high gravity anomalies traversed the whole block. Simpson and Griffiths (1982) modelled a gravity profile across the trend of the major anomalies line and interpreted the Cooper Bay dislocation as a fundamental line between an area with crust *c.* 28 km thick to the north-east and a region with an average thickness of *c.* 20 km to the south-west.

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